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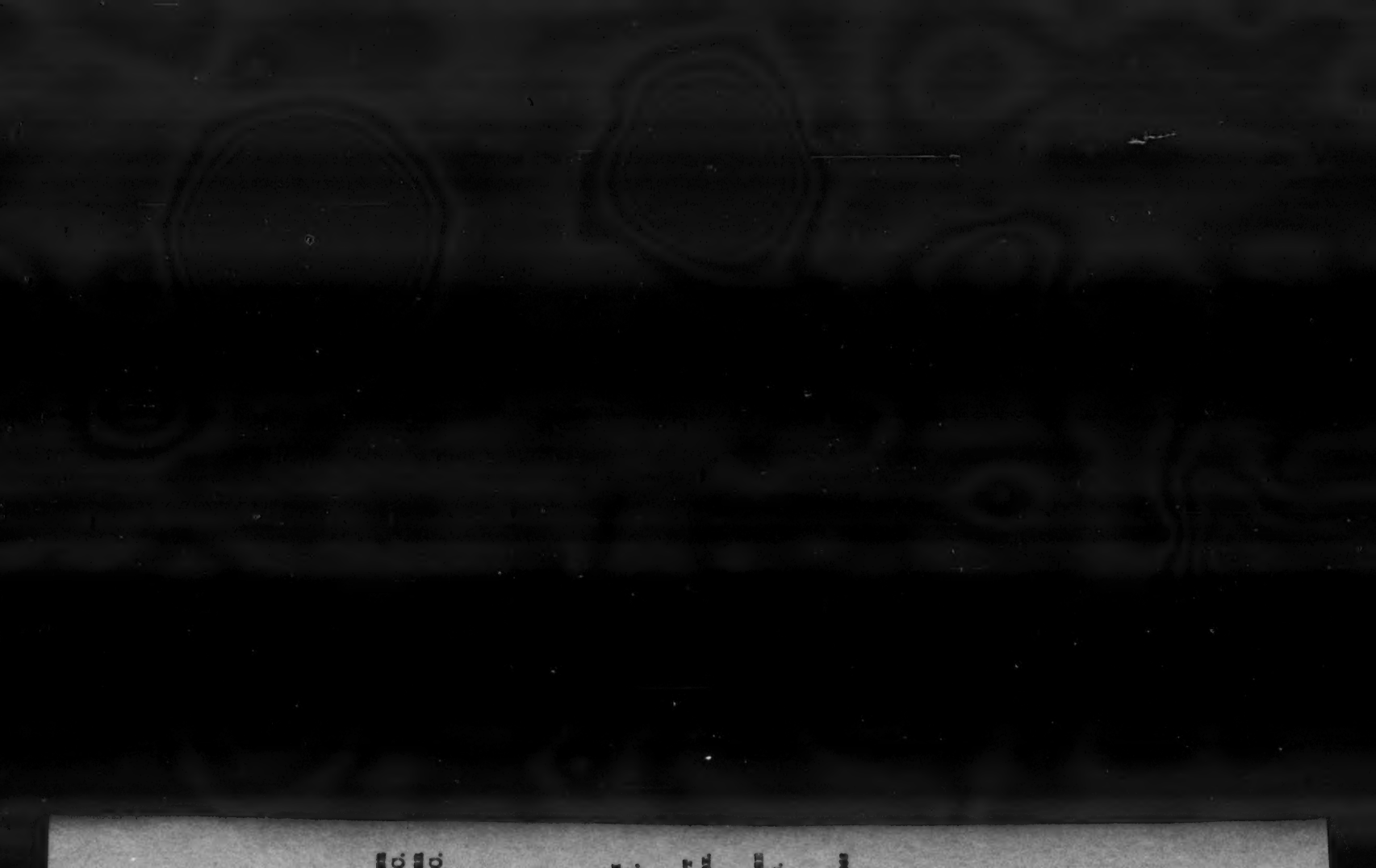

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VOLUME 64 June 1953 NUMBER 6

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 64, PP. 603-622, 13 FIGS. JUNE 1953

VERTEBRATE PALEONTOLOGY AND CONTINENTAL STRATIGRAPHY IN COLOMBIA

By R. A. Stirton

Abstract

The use of vertebrate evidence for correlation in petroleum exploration is recommended. Attention is again directed to the need for a clear understanding of the difference between time, time-rock, rock, and faunal units. Six new faunal names and one new formational name are proposed. Late Eocene, early and late Oligocene, late Miocene, Miocene or Pliocene, ? Pliocene, and Pleistocene faunal units are recognized, and the faunae are listed. The late Miocene La Venta fauna is one of the largest fossil vertebrate assemblages from South America.

The report is preliminary since most of the fossils have not been described in detail, and studies on the local geology are not complete, but there is some information on thickness of sections, sources of materials, orogenies, nature of deposition and paleoecology. A correlation chart and 11 maps and sections are included.

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Introduction

Until 1946 there were only scattered records of fossil vertebrate remains from northern South America. These were listed by Simpson (1940; 1943), but the fossils were so incomplete, the occurrences so scattered, and the observations on the geology of the localities so meager that they contributed little to our knowledge of the continental stratigraphy of that region. Yet fossil evidence from this region, and from Colombia in particular, is critical in any interpretation of the dispersal and relationships of Cenozoic vertebrate faunas on the American continents. This is true since the area lies in the most probable pathway of dispersal between the two continents. Furthermore the stratigraphic relationship of marine and continental beds can give us critical evidence on a trans-Caribbean to Gulf Coast correlation of land faunas.

Discoveries of late Eocene, early and late Oligocene, late Miocene, ?late Pliocene, and

FIGURE 1.—INDEX MAP OF COLOMBIA SHOWING AREAS WHERE FOSSIL VERTEBRATES WERE FOUND
Numbers in circles refer to numbers in upper right hand corner of detail maps.

Pleistocene assemblages (Fig. 1) have disclosed important evidence on the dispersal, evolution, and correlation of the Cenozoic vertebrate faunas in South America, though most of the specimens are still being studied and described. This then is only a preliminary report.

CORRELATION IN PETROLEUM EXPLORATION

The misconception that Vertebrate Paleontology is primarily a study in morphology and evolution, and hence is of little use in stratigraphy and correlation, still seems

prevail in the minds of many geologists. Perhaps this might be expected since the founder of modern Vertebrate Paleontology, Georges Cuvier (1769–1832), established the science on the basis of a comparative morphology in Recent animals. Furthermore he believed that each succeeding fauna, where he found it imbedded in the rocks, had died out and was buried perhaps by some catastrophic event before the next series of rocks was laid down, then a new fauna was created. Though not sharing Cuvier's ideas on catastrophism, vertebrate paleontologists of the later part of the nineteenth century in their haste to discover new and interesting fossils did little to strengthen our reputation in the field of stratigraphy.

On the other hand Invertebrate Paleontology has been thought of as the tool of stratigraphers. This undoubtedly was greatly influenced by Charles Lyell (1797–1875) and Paul G. Deshayes (1796–1896) in drawing up their Tertiary time table based on the percentage of living marine invertebrate forms represented in their Tertiary formations, a method eminently successful in pioneering correlation now nearly abandoned.

Nevertheless vertebrate paleontologists are now using their knowledge of vertebrates in correlations of stage refinement, and in certain areas early and late divisions of a stage are determined. This is possible because vertebrates, and mammals in particular, evolved rapidly and dispersed widely, even to adjacent continents, in a relatively short time. Then, too, the phyletic sequences of many groups are well known, phyletic control and stage in evolution have become accurate methods in correlation. Skeletons are not necessary; in many instances one or more teeth will suffice in making age determinations. Detailed evolution in mammals is better understood than in any class of organisms. The evolution of the horse is the best example of almost imperceptible intergradation in evolution from early Eocene to the Pleistocene. Representatives of other families, also, are nearly as well known, as exemplified in the beavers, dogs, rhinoceroses, camels, antilocaprids, and many others. Taking the evidence from the faunal assemblages as a whole, then, accurate data are available for establishing synchrony in geological and biological events locally, regionally, and even on different continents.

During the past few years the University of California has conducted expeditions to Colombia, South America, where we have dated late Eocene, early and late Oligocene, late Miocene, ?Pliocene, and Pleistocene beds. Correlations have been made with the continental sequence in Argentina (Fig. 2). Our knowledge is now sufficient to establish stage refinement in the land-laid beds in northern South America, and with continued exploration this kind of correlation can be extended throughout the continent.

One tie-in with marine beds has been made, and this can be done in many other areas along the Caribbean Coast. Every effort is being made, despite limited funds, to carry out detailed stratigraphic studies in the areas where fossils have been found. These include thickness of sections, sources of materials, orogenies, nature of deposition, and paleoecology.

Along the Andes from Venezuela to Argentina, there are many thousands of feet of continental beds representing every epoch of the Cenozoic. In many places these sedimentary units overlie possible reservoirs of petroleum. During the past 20 years, millions of dollars have been spent on research and on field geology in search for new oil fields, yet one of the most effective tools in correlation, Vertebrate Paleontology, has been almost totally neglected. Perhaps this is because vertebrate remains are usually more difficult to locate in the field than are invertebrates, but trained field men find them, as demonstrated by our work in Colombia. It is true that some beds are almost devoid of fossil remains, but this is the case for both continental and marine beds. With adequate funds and facilities more important stratigraphic studies could be continued profitably in some of the more critical areas.

Stratigraphic and Faunal Nomenclature

The most distressing experiences in assembling the information for this report were in trying to find satisfactory terminology for the rock units containing the vertebrate remains. For the most part there has been little or no

	COLOMBIA	ARGENTINA	UNITED STATES	EUROPE
	FAUNAS	'STAGES'	PROVINCIAL AGES	STAGES
PLEISTOCENE	Cátedras	Pampean Ensenadan	Rancholabrean Irvingtonian	
PLIOCENE	Cocha Verde ?	Uquian Chapadmalalan Montehermosan Tunuyanian Huayquerian Mesopotamian	Blancan Hemphillian Clarendonian	Astian-Villafranchian Pontian Sarmatian
MIOCENE	Peñata LaVenta,Carmen de Apicalá	Chasicoan Friasian Santacrucian	Barstovian Hemingfordian Arikareean	Vindobonian Burdigalian Aquitanian
OLIGOCENE	Coyaima Chaparral,Peneyita	Colhuehuapian Deseadan	Whitneyan Orellan Chadronian	Chattian Stampian Sannoisian
EOCENE	Tama	Mustersan Casamayoran	Duchesnian Uintan Bridgerian Wasatchian	Ludian Aversian Lutetian Cuisian Sparnacian
PALEOCENE		Riochican	Clarkforkian Tiffanian Torrejonian Dragonian Puercan	Thanetian

FIGURE 2.—TENTATIVE CORRELATION CHART

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differentiation between *time*, *time-rock* and *rock* units (*see* Schenck and Muller, 1941), a discipline which no longer can be ignored in stratigraphic and historical geology. Many previous writers have confused groups with series, formational names have been given stage connotation, and little regard has been given to priority in nomenclature. In one instance three authors working in one area have applied three different sets of names to the same stratigraphic units.

One of the most important projects for future studies on the stratigraphy in Colombia, then, is to revisit the type localities of the named stratigraphic units, map the areas, draw the stratigraphic sections, and attempt to understand the sense of the original author. A name like Honda need not be abandoned because some authors failed to use it in the sense of the original author, nor need it be discarded if it is later found to be referable to a higher category in rock terminology—*i.e.*, group instead of formation. It is further recommended that the law of priority and typology in stratigraphic nomenclature (Stirton, 1951b) be applied if the confusion of names currently in use in northern South America is to be clarified.

To avoid confusion, faunal names (Stirton, 1936; 1939) instead of stratigraphic terms have been used in this report, though an effort has been made to apply the correct rock unit name. A single extinct species or a tooth of an individual may be our sole representation of a distinct faunal unit. Since a fauna is all the animals in any given area at any given time, it would be impossible to have total representation in any fossil assemblage. The term faunule, then, as it is sometimes applied to a meager assemblage, is considered ambiguous. As in taxonomy, additional evidence may necessitate synonymizing certain faunal names (La Venta and Carmen de Apicalá), and one name (La Venta or Tama) as it now stands may eventually refer to two or more faunas. This is only normal procedure in the increment of our knowledge and need not be confusing or misleading.

The necessity for using faunal units as distinct from rock units is well exemplified in the Coyaima (late Oligocene) and the La Venta (late Miocene) faunas which are found in one stratigraphic unit, the Honda. On the other hand the Chaparral and the Peneyita may be an example of the same fauna occurring in two distinct stratigraphic units.

Faunal and temporal units are likewise treated separately. A grassland fauna in the Llanos and a forest fauna in the lower Magdalena Valley, though contemporary, may show very little in common. This is similar to the ecofacies problems currently under consideration by invertebrate paleontologists. New faunal and stratigraphic names appear in **boldface** type.

It perhaps would be advisable to set up some marine stage terminology for South America, but for the continental sequence Simpson's (1940) terminology is adequate at least for the present.

Acknowledgments

Our field work in Colombia was initiated in 1944 under a Guggenheim fellowship. This was a co-operative project of the Servicio Geológico Nacional de Colombia and the Museum of Paleontology, University of California. It was partly financed by the International Petroleum Company, the Shell Oil Company, the Socony Vacuum Oil Company, the Richmond Oil Company, the Texas Petroleum Company, and the Colombian Gulf Company.[1] Later expeditions were also sponsored by the Associates in Tropical Biogeography, University of California. Thanks are due Drs. Donald E. Savage, Wann Langston, Jr., and Robert W. Fields for their suggestions in some of these preliminary identifications. Owen J. Poe prepared the illustrations. The generous support and keen appreciation of these projects by Dr. Enrique Hubach, Director, Servicio Geológico Nacional, has made this work both pleasant and profitable.

Vertebrate Paleontology and Stratigraphy

Eocene

TAMA fauna: Late Eocene; Mugrosa formation

Class—TELEOSTOMI
Order—PERCIFORMES
Family—?Sparidae; numerous teeth

[1] Full acknowledgments given to all participants in Stirton (1951a) not repeated here.

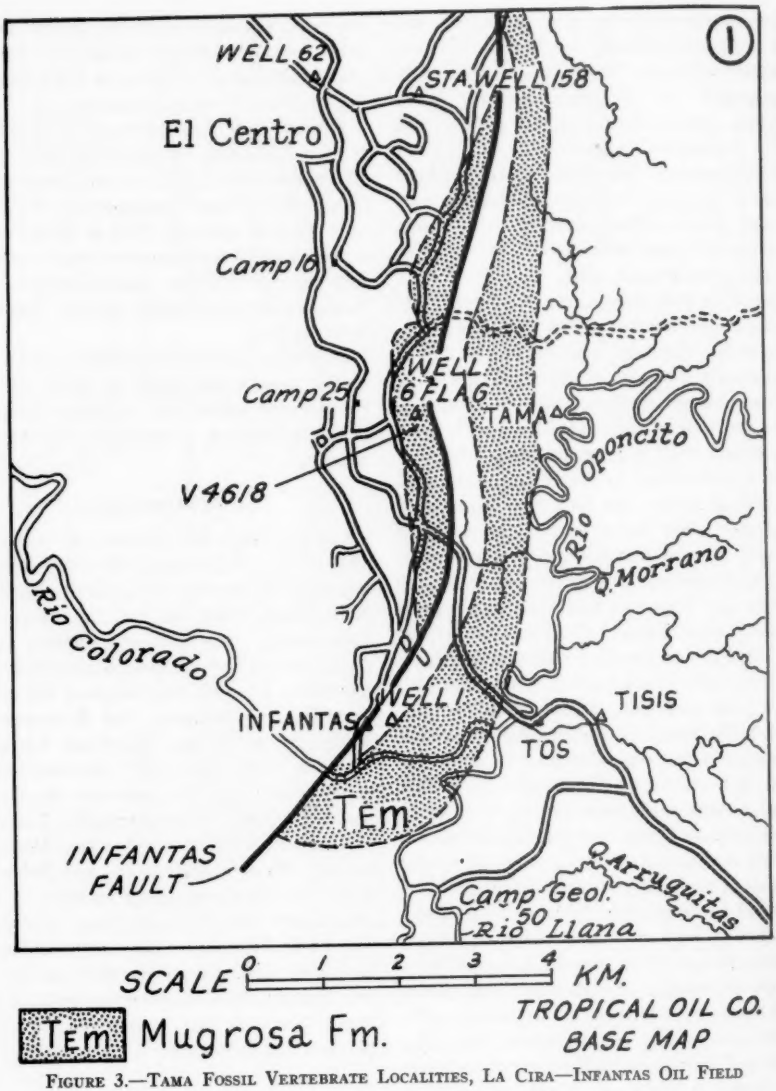

FIGURE 3.—TAMA FOSSIL VERTEBRATE LOCALITIES, LA CIRA—INFANTAS OIL FIELD

Class—REPTILIA
Order—CHELONIA; fragments of plastrons and carapaces
Order—SEBECOSUCHIA
Family—Sebecidae. *Sebecus*; part of tooth
Order—EUSUCHIA
Family—Crocodylidae; two slender teeth with corrugate surfaces
Class—MAMMALIA
Order—ASTRAPOTHERIA
Family—Astrapotheriidae
Subfamily—Albertogaudryinae; lower molar

Fossil vertebrates discovered in the Mugrosa formation are the only Eocene vertebrates known from Colombia. These occur in the La Cira-Las Infantas oil fields in Santander (Fig. 3). Two kinds of crocodilians, turtles, and a molluscivorous fish were found in the two invertebrate horizons along the bank of the Rio Llana at Camp Waldo (Loc. V 4620). This evidence seems to indicate a fresh-water or possibly partly brackish-water environment.

The only evidence of a fossil mammal is an astrapothere lower molar. The specimen was found in a grayish lens about half way between wells

and 375, type locality of Tama fauna (Loc. V 4618). This is below the invertebrate horizons but still in zone B (J. L. Anderson, 1945, p. 1094). Though I cannot identify the genus, no astrapothere in this stage of evolution has been found in the early Eocene of South America, and apparently nothing as primitive occurs in the early Oligocene.

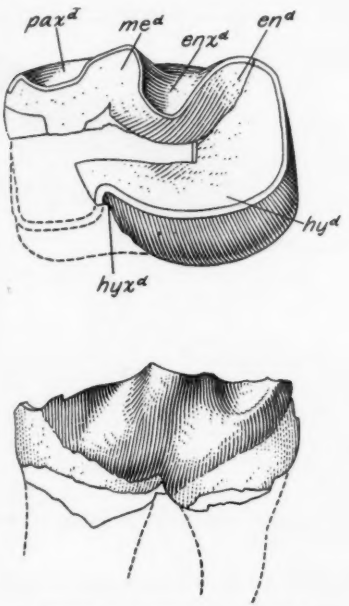

FIGURE 4.—ALBERTOGAUDRYINAE

Left M_1 or M_2, occlusal and lingual views, natural size. Tama fauna, Mugrosa formation, late Eocene.

DESCRIPTION: M_1 or M_2, heavily worn (Fig. 4); anterolabial corner broken and lost. Brachyodont; ectolophid not straight, paralophid continous from protoconid to anterior lingual corner, protoconid labial in position from line connecting inner edges of metaconid and entoconid, with strong lingual direction opposite hypoconid, hypoflexid evidently well developed in upper half of crown but fades out toward base; entoflexid widely U-shaped, bottom with verticolabial direction; entoconid area larger than metaconid area; no cingulum on posterior moiety; anterolingual cingulum extending from paraconid down to base of paraflexid. Length along inner side—42.0 mm; width across hypoconid 29.0 mm. Anterior moiety narrower.

This tooth seems clearly referable to the Astrapotheriidae, and evidently referable to the subfamily Albertogaudryinae. It is relatively lower-crowned and wider than a molar of *Astraponotus* (Am. Mus. No. 29449) from the Musters formation in Argentina. The *Astraponotus* tooth differs also in its prominent cingulum and in the anterior tip of the trigonid which is extended less lingually. The exact relationship of the Tama tooth to *Albertogaudrya* from the early Eocene and to *Astraponotus* from the middle Eocene is not clear. It may be a direct descendant from *Albertogaudrya* but it is larger and has no isolated lower cusp. There probably are other differences from the early Eocene genus.

CORRELATION: The Tama astropathere differs from the Oligocene and Miocene genera in its lower crown and lack of alignment of the lingual cusps. It may represent a distinct genus, but it seems advisable not to confound taxonomists in the future with such an inadequate type. At present it will serve in correlation problems, particularly when other remains are found in Eocene deposits in northern South America.

Oligocene

CHAPARRAL fauna: Early Oligocene.
"Tuné" formation
(Stirton, 1946b)

Class—TELEOSTOMI; vertebrae and other bone fragments
Class—AMPHIBIA
  Order—SALIENTIA; femur
Class—REPTILIA
  Order—CHELONIA; plaques and fragments of large chelonians
  Order—EUSUCHIA
    Family—Alligatoridae; plaques and teeth
    Family—Crocodylidae; teeth
Class—MAMMALIA
  Order—EDENTATA
    Superfamily—Megalonychoidea; distal end of metapodial and part of cheek tooth
  Order—INCERTAE SEDIS
    Family—Incertae sedis. *Lophiodolodus chaparralensis*; holotype; M_2
  Order—LITOPTERNA
    Family—Proterotheriidae. *Protheosodon*; inner half of second upper molar
  Order-NOTOUNGULATA
    Family—Toxodontidae. *Proadinotherium*; poorly preserved upper molar
  Order—ASTRAPOTHERIA
    Family—Astrapotheriidae; six isolated cheek teeth; tip of upper incisor (the latter was previously referred to "?toxodont")

The Chaparral faunal locality (Fig. 5) was found in the "Tune" formation (Texas Petroleum Company terminology, here included in the Gualanday group), which is reddish or gray and composed of claystones, shales, sandstones, and conglomerates that were deposited on a low alluvial plain. The beds are strongly folded and faulted. They overlie formations of probable Eocene and Paleocene age which also appear to be continental deposits (Fig. 6). Conformably above the "Tune" are gray sandstones and claystones including some red members. This is the base of the Honda group. (*See* discussion under La Venta.) An Eocene age for the underlying "Carrasposa" (Texas Petroleum Company terminology, here included in the Gualanday group) formation, could not be confirmed due to inadequacy of the vertebrate materials discovered (Locs. V4409 and V4410).

Observations and fossil discoveries were made south of Rio Tetuán 6 kilometers northeast of Chaparral on and adjacent to the San Jose anticline and along the carretera (highway) from Coyaima at kilometer 39, Tolima.

Fresh-water invertebrates found in the same

horizon as the main vertebrate locality V4406 have been referred tentatively to the La Cira fauna as typified in the Sogamosa River area of the middle Magdalena region (Pilsbury and Olson, 1935). This evidence together with that of abundant caiman, crocodilian, chelonian, and fish remains is indicative

does not agree in detail with the described tooth of this genus from the Deseado, it seems clearly more advanced than those from the Musters (middle Eocene) and more primitive than those from the Colhué-Huapí (late Oligocene). The species of *Proadinotherium* range through the Oligocene, and

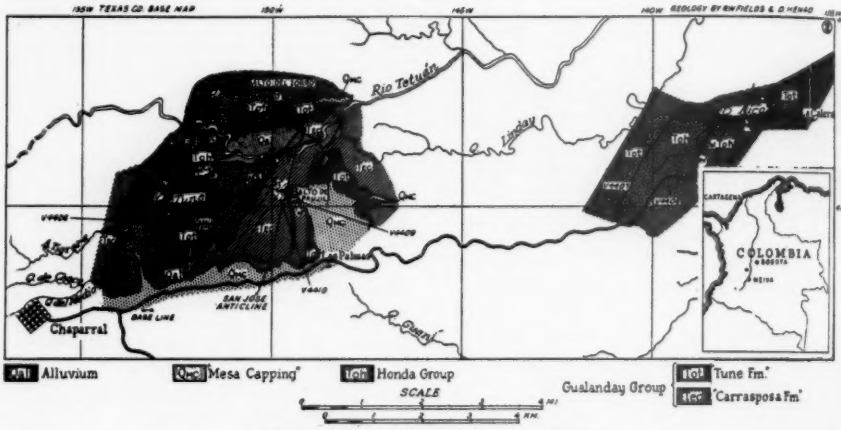

FIGURE 5.—CHAPARRAL REGION, TOLIMA
See section A-B-C-D in Figure 6.

of lakes and/or pools in stream courses over a lowland alluvial area. The lowland must have been well populated with mammals of which we have five orders, probably mostly browsing and partly open forest and amphibious types. Since astrapotheres are more common than the other mammals this may support the idea of their being somewhat amphibious. The ground sloth, the litoptern *Protheosodon*, and the toxodont *Proadinotherium* may have approached this reptile-infested pool only for a drink or to cross one of its source streams. *Lophiodolodus* on the other hand may have been a forest animal, and a relic in equatorial latitudes. The bodies of all such animals that chanced to be caught in these waters could have been dragged out into the lake, torn apart, and for the most part eaten by swarms of caimans and crocodiles. In this operation some parts sank to the bottom where they settled into the sandy mud. Over 300 teeth of these predaceous reptiles were collected in a very limited area. Coprolites of these creatures were equally abundant. The limited exposure on this dip slope where the fossils were found is made up primarily of the residue that settled to the bottom in this tropical Oligocene pool. Every fossil was fragmentary.

CORRELATION: The best evidence for an age determination of the Chaparral fauna is from the mammal remains. These materials have been compared with the assemblages from Argentina, and the most marked resemblance is with the Deseado (early Oligocene).

Perhaps the most indicative specimen in the Chaparral assemblage is the inner half of a *Protheosodon* tooth, M^2. Though as expected the tooth

evidence from Chaparral is insufficient to make a closer correlation.

PENEYITA fauna: Early Oligocene. ? formation

Class—MAMMALIA
Order—RODENTIA
Family—Erethizontidae. *Eosteiromys*; part of a maxillary with P^4.

This specimen was found in a calcareous marly claystone—"at the confluence of the Rio Peneyita with the Rio Caquetá" (Renz),—Caquetá (Fig. 7). The writer has not seen the locality and has no additional information on the geology of the region. This is our only vertebrate record east of the Andes in Colombia. Its nearest correlative in the Magdalena valley is the Chaparral.

At another locality in this region part of a large ?caiman scapula was found in "Tertiary clays and ferruginous sandstones with *Unio* and fresh-water fossils, at the Rio Guayabero five miles below the mouth of the Rio Heroru." (A. A. Olsson and Donald Macgregor). We cannot correlate this with the ?caiman.

COYAIMA fauna: Late Oligocene.
Honda group

Class—TELEOSTOMI
Order—CYPRINIFORMES; large catfish spine
Order—PERCIFORMES; one cheek tooth
Class—REPTILIA
Order—CHELONIA; parts of plastrons and carapaces

Order—SAURIA; mandibular fragment of large lizard or iguana
Order—EUSUCHIA
Family—Alligatoridae; plaques and teeth
Family—Crocodylidae; plaques and teeth
Family—Gavialidae; parts of two craniums

type locality just above the city of Honda along the Magdalena River, nor is there as much grayish-green claystone as can be found in the badlands near the village of Villavieja farther up the Magdalena. The stratigraphic terminology will be discussed under La Venta fauna.

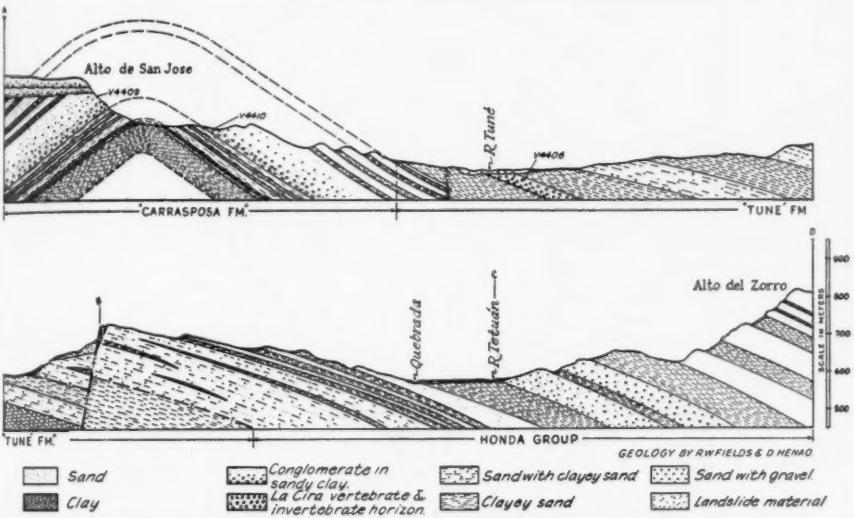

FIGURE 6.—ALTO SAN JOSE—ALTO DEL ZORRO SECTION NEAR CHAPARRAL, TOLIMA
See areal geology in Figure 5. The upper and lower sections are continuous.

Class—MAMMALIA
Order—EDENTATA
Family—Megalonychidae; fibula and tooth
Family—Dasypodidae; two scutes of small form, and numerous scutes of large chlamythere
Order—RODENTIA
Family—Dinomyidae; upper molar and two incisors related to *Scleromys*
Order—LITOPTERNA; lower end of calcaneum
Order—NOTOUNGULATA
Family—Toxodontidae; lower milk tooth
Family—Leontiniidae; two upper molars in United States National Museum said to have come from this area (information from Dr. C. L. Gazin).
Family—Interatheriidae. *Cochilius*; maxillaries and palate with associated mandibles
Order—ASTRAPOTHERIA
Family—Astrapotheriidae; palate with cheek teeth and isolated teeth

The Coyaima faunal remains were discovered in gray ferruginous sandstones near the village of Coyaima adjacent to the Castilla-Ataco carretera (Fig. 8). These have been considered part of the Honda group with abundant hornblende and slightly less black opaque minerals as a characteristic feature (Butler, 1942). The beds here are not composed of such coarse clastics as at the

The fossil-bearing beds at Coyaima with numerous quartz and andesite pebbles seem to rest conformably on red claystones and reddish conglomerates. This relationship is exposed west of the Coyaima-Ataco carretera between kilometers 6 and 8. A similar relationship is shown about 5 or 6 kilometers beyond La Calera in a synclinal remnant of these gray sandstones on the Coyaima-Chaparral carretera, where some unidentifiable vertebrate remains were found in the underlying sedimentary rocks; thus we were unable to establish the age of the underlying red claystones and reddish conglomerates. Possibly they are part of the "Tuné" formation, though the upper part of that formation may contain middle Oligocene fossils.

Fossils are scattered throughout the gray sandstones, particularly turtle, caiman, and crocodilians. Mammal bones and teeth are much less common. Evidently the area was still a nearly flat alluvial plain, frequently flooded when large logs and trunks of trees were floated in and buried in the sand. The remains of some of these logs can be seen west of the Coyaima-Ataco carretera near locality V4417. One was taken to the museum in Bogotá.

The most productive single locality is a small erosional remnant (Loc. V4411) about 12 meters long and 1 to 3 meters wide. This is slightly more than a kilometer from Coyaima and on the south side of the Coyaima-Castilla carretera. Here red claystones are overlain by nearly a meter of coarse friable sand, well-rounded gravel, small clay balls,

and tiny concretions of calcium carbonate. This evidently represents the remnant of a shallow pool of limited extent where the scattered bones and teeth of fish, caimans, crocodilians, a large iguana-like creature and a few mammalian remains sank to the bottom. A few internal casts of fresh-water gastropods were discovered, and coprolites were numerous. The mammal bones included plaques of a small armadillo, part of the calcaneum of a small litoptern, and the maxillaries, palate, and mandibles of a small interathere, *Cochilius*. Part of a mandible of another *Cochilius* was also picked up on the surface. The fossils occur as float on the slope and come from the cap rock.

FIGURE 7.—PART OF CAQUETA—PUTAMAYO REGION NEAR PERUVIAN BORDER, SHOWING PENEYITA LOCALITY

CORRELATION: The Coyaima interathere, *Cochilius*, is closely related to species from the Oligocene of Argentina. It is slightly larger than *C. volans* Ameghino, but both *C. pendens* Ameghino and *C. columnifer* Ameghino from the Deseado are larger. The genus *Cochilius* evidently ranges throughout the Oligocene in Argentina. On the other hand the scleromyid rodents, of which we have an upper molar and parts of two incisors, do not range below the early Miocene in Argentina, though they may eventually turn up in the Colhuehuapian. The most logical conclusion then is that we are dealing with a late Oligocene assemblage at Coyaima. Until more critical parts of the other mammals have been found they will not help in this more refined correlation.

Miocene

LA VENTA fauna: Late Miocene. Honda group (Cabrera, 1929; Kraglievich, 1928; Mook, 1941; Reinhart, 1951; Royo y Gómez, 1942a, 1942b, 1942c; 1945; Savage, 1951a; 1951b; Simpson, 1940; 1943; Stirton, 1951a; Stirton and Savage, 1951).

Class—DIPNOI
  Order—LEPIDOSIRENIFORMES
    Family—LEPIDOSIRENIDAE; numerous teeth
Class—TELEOSTOMI
  Order—CYPRINIFORMES
    Suborder—SILURI; catfish, numerous spines

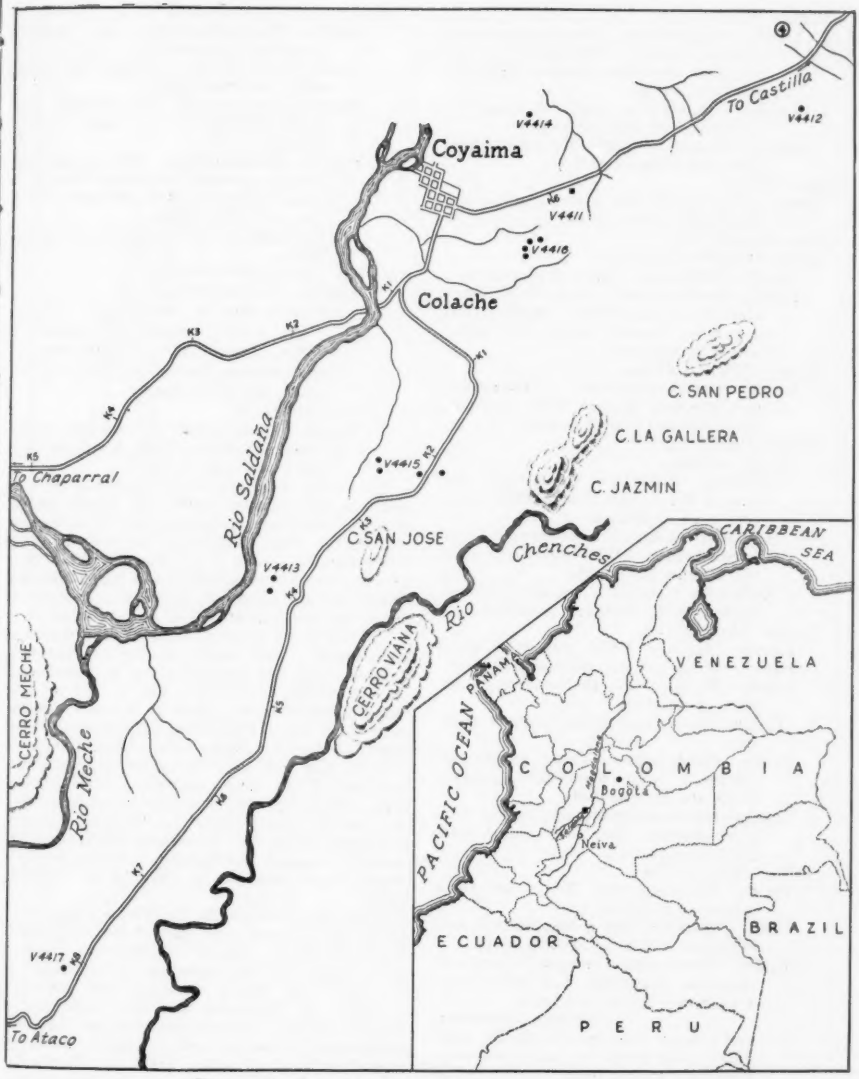

FIGURE 8.—COYAIMA FOSSIL VERTEBRATE LOCALITIES

Order—PERCIFORMES
Families not identified
Class—AMPHIBIA
Order—SALIENTIA
Family—Leptodactylidae; parts of skeleton
Class—REPTILIA
Order—CHELONIA
Family—Chelyidae; plastron and carapace
Family—Pelomedusidae; plastron and carapace
Family—Emydidae; miscellaneous parts
Family—Testudinidae; plastron and carapace
Order—SAURIA
Family—Teiidae; mandibles
Probably other families
Order—SERPENTES; vertebrae
Order—SEBECOSUCHIA
Family—Sebecidae. *Sebecus*; part of mandible and isolated teeth

Order—EUSUCHIA
Family—Alligatoridae. *Brachygnathosuchus neivensis* (Mook), 1941; holotype, and craniums, mandibles, isolated teeth and parts of skeletons
Family—Crocodylidae; parts of craniums, mandibles and isolated teeth
Family—Gavialidae; parts of craniums, mandibles and isolated teeth
Class—AVES
Order—GALLIFORMES; roof of cranium and other unidentifiable bones
Class—MAMMALIA
Order—MARSUPIALIA
Family—Didelphidae; parts of mandibles
Family—Borhyaenidae. *Lycopsis* skeleton. *Cladosictis*, fragmentary cranium and mandible. Another larger form represented by fragments
Order—CHIROPTERA
Family—Phyllostomatidae. *Notonycteris magdalenensis* Savage, 1951; holotype; mandibles and part of skeleton
Order—PRIMATES
Family—Cebidae
Subfamily—Pithecinae. *Cebupithecia sarmientoi* Stirton and Savage, 1951;[2] holotype; part of skeleton
Subfamily—Alouattinae. *Homunculus tatacoensis* Stirton, 1951a; holotype; mandible. *Homunculus*, tooth
Subfamily—Cebinae. *Neosaimiri fieldsi* Stirton, 1951a; holotype; mandible
Order—EDENTATA
Family—Megalonychidae; parts of skeletons
Family—Megatheriidae; parts of skeletons
Family—Mylodontidae. Perhaps four genera, parts of skeletons
Family—Myrmecophagidae; parts of skeletons
Family—Dasypodidae. Three genera including a large chlamythere, plaques, teeth, and skeletal parts
Family—Glyptodontidae; some nearly complete skeletons, isolated craniums, mandibles, and plaques
Order—RODENTIA
Family—Erethizontidae. Two cheekteeth related to *Eosteiromys*
Family—Caviidae. One genus, parts of two skeletons, and other isolated parts.
Family—Dinomyidae. *Olenopsis*;[3] parts of craniums, mandibles, and body skeletons; numerous isolated teeth
Scleromys schürmanni
and a larger *Scleromys*; craniums, mandibles, and parts of body skeletons; numerous isolated teeth
Order—CONDYLARTHRA
Family—Didolodontidae; part of mandible
Order—LITOPTERNA
Family—Macraucheniidae; teeth and foot bones
Family—Proterotheriidae; teeth and parts of of body skeleton
Order—NOTOUNGULATA
Family—Henricosborniidae; M_1 (possibly secondarily deposited)
Family—Leontiniidae; cranium, mandibles, teeth, and limb bones
Family—Toxodontidae. Two genera; craniums, mandibles, and parts of body skeleton
Family—Interatheriidae. One genus and species; numerous skeletons varying in completeness, isolated craniums, mandibles, and other bones
Family—Hegetotheriidae, cranium and mandibles
Order—ASTRAPOTHERIA
Family—Astrapotheriidae
Xenastrapotherium kraglievichi Cabrera, 1929; palate, maxillaries, mandibles, tusks, isolated teeth, and limb bones
Astrapotherium, mandibles, tusk, isolated teeth, and limb bones
Order—SIRENIA
Family—Trichechidae; *Potamosiren magdalenensis* Reinhart, 1951, holotype, mandible and isolated M_1.

The La Venta fauna is one of the largest and most complete late Miocene vertebrate faunas known, and it is even more interesting since it occurs near the equator. Here were found not only numerous new genera but relics of families thought to have become extinct in earlier epochs in higher latitudes. There also are numerous forms of wide range in South America that are useful in correlation.

The La Venta fossils were found in a section of approximately 700 meters including sandstones, conglomerates, and claystones, but no volcanic ash (Stirton, 1951a, p. 316–317), and to date we have found no evolutionary difference in fossils from the base to the top of this section. The section studied extends from Cerro Gordo to and beyond Villavieja (Fig. 9). The beds also occur in an anticlinal structure south of Aipe on the west side of the Magdalena. East of Villavieja and on south toward Neiva they are overlain conformably by the predominately tuffaceous "Gigante" formation (Richmond Petroleum Company terminology), and this in turn is capped unconformably by the "Mesa" conglomerates.

The beds in which the La Venta fauna occurs have been referred to as part of Hettner's (1892) Honda formation. Much confusion has arisen in determining the stratigraphic relationships of Hettner's type Honda to other rock units—for example, the Barzalosa and the Gualanday of Scheibe (1922). Some geologists feel that Barzalosa is the correct name for these rock units, and others would use Real (Wheeler, 1935), but Honda has priority. Butler (1942), who used the term Honda, directed attention to these and many other difficulties in applying stratigraphic terminology in the Magdalena Basin. His broadening of the terminology is commendable, but it is preferable to use Honda *group*, a rock term, instead of *series*, a time-rock term, at least until detailed reports on the stratigraphy of the middle and upper Magdalena basins are published.

Our vertebrate evidence has disclosed that the gray sandstones and claystones referred to the Honda near Coyaima are late Oligocene, and the rocks in the La Venta badlands are late Miocene. Un-

[2] Separates issued February 28, 1951.

[3] The placing of these genera in the Dinomyidae must be credited to Robert W. Fields who is writing his Ph.D. dissertation on the hystricomorph rodents in these faunas.

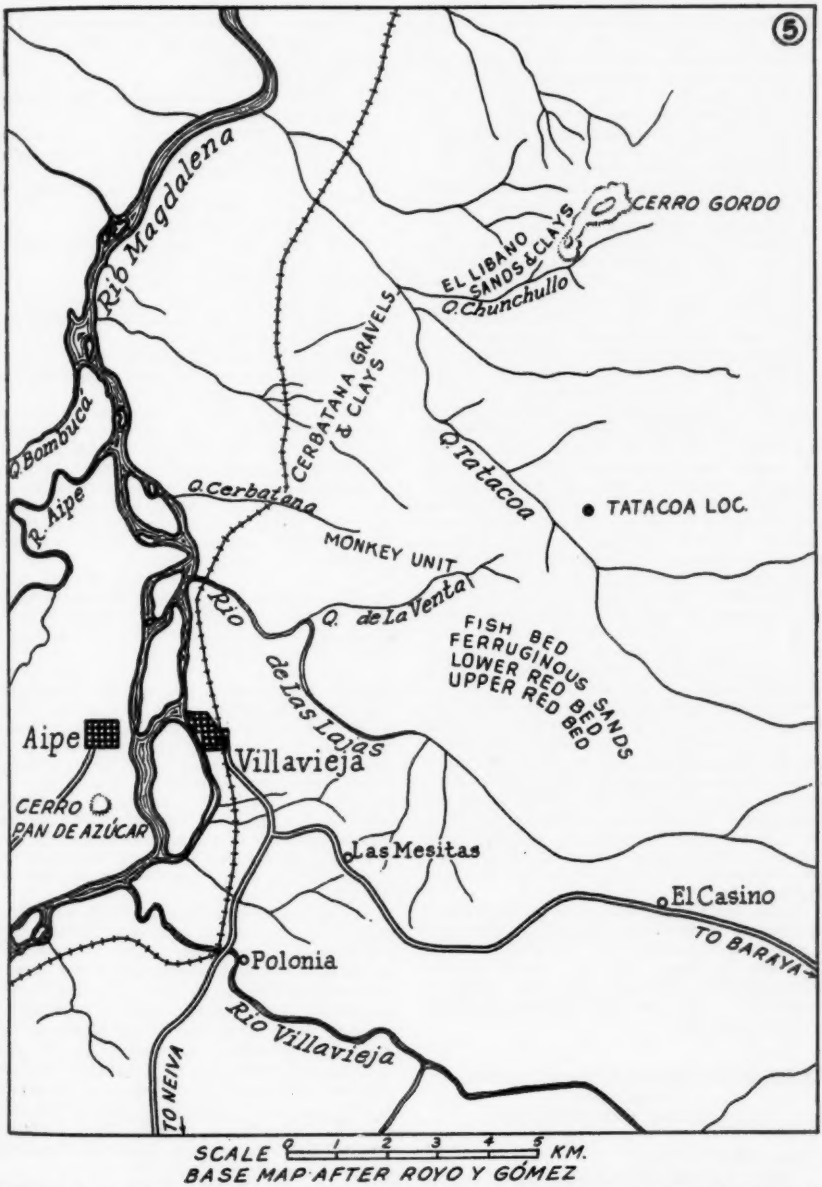

FIGURE 9.—LA VENTA LATE MIOCENE FOSSIL VERTEBRATE LOCALITIES BETWEEN CERRO GORDO AND LAS MESITAS

The areal geology of this area will appear in a paper by R. W. Fields.

fortunately no fossils have been found in the type area of the Honda. Quite possibly these beds, though widely separated in time and space in the Magdalena valley, could properly be referred to as

a formation. Yet the writer believes this is doubtful and prefers to reserve judgment on this and refer to the vertebrate materials as faunal units.

Evidence indicates that the Honda group was laid down as rather widespread alluvial-fan and flood-plain deposits and may have extended, as Butler (1942) suggests, over part of the area now occupied by the Cordillera Oriental. Evidently there were periods of rather heavy rainfall and/or local orogenies resulting in the deposition of 400 or 500 meters of sands and gravels. These were preceded and followed by widespread accumulations of sandy silts and muds in lowland depressions.

Present evidence does not permit an accurate restoration of the environment throughout the area, for it may have varied in different places or even in the same places at different times. Nevertheless there must have been wide stretches of savanna in the La Venta, Rio Tatacoa, and Cerro Gordo areas with heavy forests near the streams. Oxbows developed and, isolated by meandering streams, became mud traps for terrestrial animals as the water evaporated during the dry seasons. Interathere and glyptodont remains were abundant where the animals bogged down in the mudflats. The streams and pools were populated with fresh-water gastropods, crustaceans, fish, crocodilians, chelonians, and natatorial mammals. The great number of kinds and of individuals indicates keen competition for survival. The only obviously arboreal animals are the monkeys, though some of the smaller sloths may have been at least partly arboreal.

CORRELATION: The full significance of this large faunal assemblage cannot be realized until each group has been studied in detail and described. There is enough evidence now available on the rodents and interatheres to recognize a late Miocene age for the fauna.

The rodents that Fields has referred to, the Dinomyidae, are clearly more advanced than the few representatives in the Santacrucian (early Miocene) and not so progressive as their descendants in the Mesopotamian (early Pliocene). Unfortunately the group to which the La Venta interathere belongs is not represented in the Santa Cruz formation. Nevertheless its stage of evolution when compared with *Cochilius* from the Colhuehuapian (late Oligocene) is too advanced for an early Miocene form.

The La Venta would seem to be nearer in age to the Friasian of Argentina than to the other ages given by Simpson (1940), but the fossils from the Río Frías are meager and not clearly allocated stratigraphically. Problems like this could be clarified with a little field work along the Río Frías and in adjacent areas.

CARMEN DE APICALÁ fauna: Late Miocene.
Honda group
(Royo y Gómez, 1945; Stehlin, 1939)

Class—TELEOSTOMI
  Order—CYPRINIFORMES
    Suborder—SILURI; catfish spines
  Order—PERCIFORMES; teeth and spines
Class—REPTILIA
  Order—CHELONIA
    Family—Pelomedusidae; parts of plastrons and carapaces
  Order—EUSUCHIA
    Family—Alligatoridae; isolated teeth
    Family—Crocodylidae; isolated teeth
Class—MAMMALIA
  Order—EDENTATA
    Family—?Myrmecophagidae; two phalanges
  Order—RODENTIA
    Family—Dinomyidae; *Scleromys schiirmanni* Stehlin, 1939; holotype, maxillary: paratypes, other isolated teeth
    A larger *Scleromys*, fragments of teeth
  Order—LITOPTERNA
    Family—Proterotheriidae; distal end of femur and navicular
  Order—NOTOUNGULATA
    Family—Leontiniidae; tooth fragments
    Family—Toxodontidae; lower molar and tooth fragments
  Order—ASTRAPOTHERIA
    Family—Astrapotheriidae; humerus and tooth fragments

The Carmen de Apicalá fauna was found in the Melgar basin 20 kilometers southeast of Girardot, Tolima (Fig. 10). The lithology is much like that of the Honda and seems clearly referable to that group. The formation is about 112 meters thick in the south end of the basin and perhaps one-third thicker to the north along the Carmen-Girardot carretera. The basin is an asymmetrical syncline with north-northwest-south-southeast axis.

The members of the formation are gray ferruginous sandstones and gray to red claystones, with some rather thin conglomeratic members with tiny clay balls and small calcareous concretions. None of the concretions or gravels in the concretionary units are more than an inch in diameter, and most of them are much smaller. Two of these conglomeratic members in the south part of the syncline were the main fossil-bearing units, though chelonian, crocodilian, and astrapothere remains as well as the toxodont molar, fossil wood, and poorly preserved leaves were found elsewhere in the section. The conspicuous light-gray, fine-grained sandstones north of the village are cross-bedded. The fine-grained character of the sandstones and the small size of the pebbles seem to indicate that at the time of deposition of these beds the area had fairly low relief.

As in the Coyaima area, probably these light-colored beds are conformable upon the underlying red beds, here tentatively called the Gualanday group. In many places near the contact the members appeared so conformable and so similar in color and texture that it was impossible to know whether we were on the Gualanday or the Honda. Farther down in the section the Gualanday was more reddish, and there were thick members of coarse well-rounded conglomerate, coarse- to fine-grained sandstones, and red claystones with grayish-green streaks and splotches. Possibly two formations separate the Cretaceous and the Honda, but a good contact was not seen.

CORRELATION: The rodent *Scleromys schiirmanni* as well as the larger *Scleromys*, to be described by Robert W. Fields, also found in the La Venta, is good evidence for a synchronous correlation with that fauna, though no evidence was found of the large *Olenopsis* so abundant in the La Venta. The

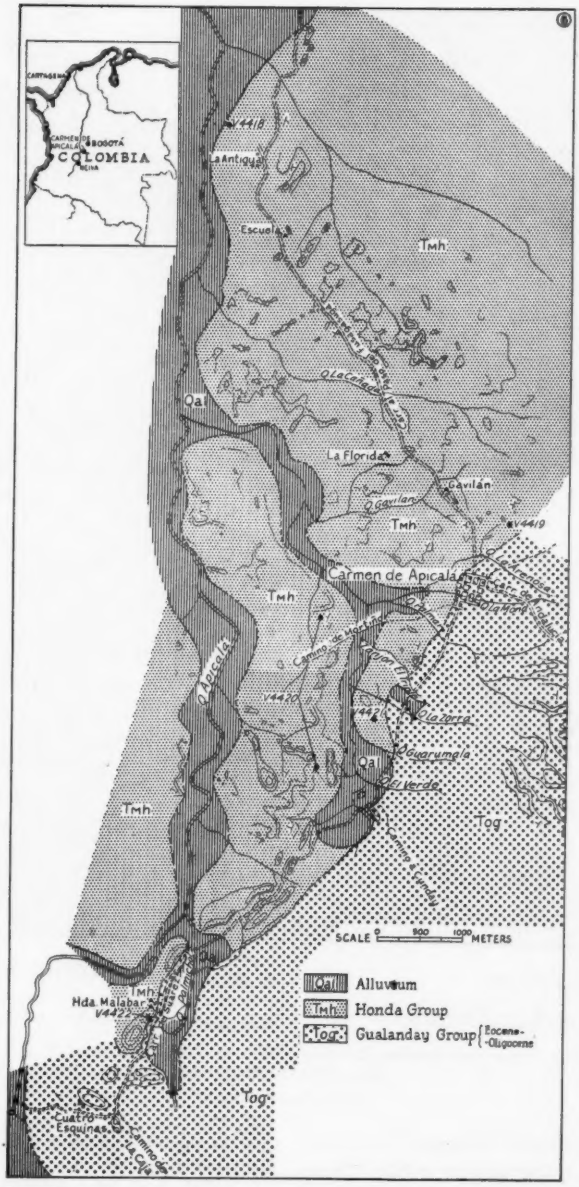

FIGURE 10.—CARMEN DE APICALA LATE MIOCENE FOSSIL VERTEBRATE LOCALITIES, TOLIMA
After Royo y Gómez.

toxodont lower molar is in a Miocene stage of evolution but the remainder of the fossils are too incomplete to use in such refined correlation. Available evidence dates the Carmen de Apicalá fauna as late Miocene and either the same or a near contemporary of the La Venta fauna.

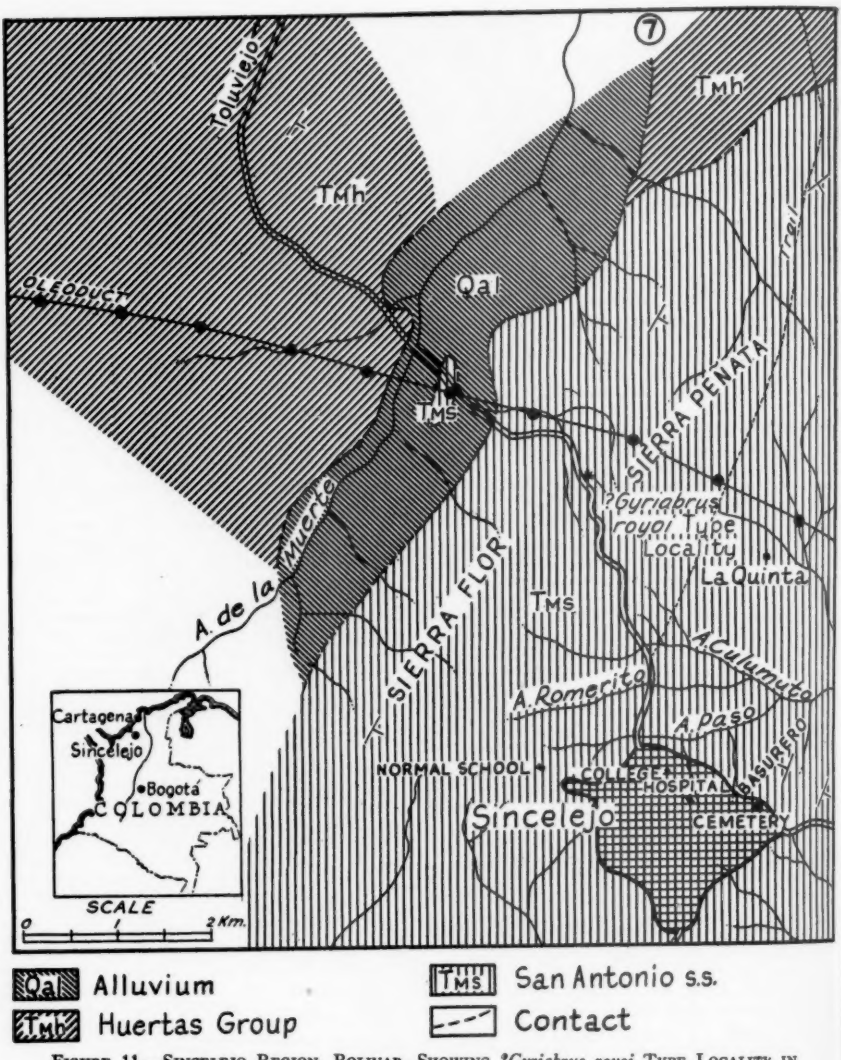

FIGURE 11.—SINCELEJO REGION, BOLIVAR, SHOWING *?Gyriabrus royoi* TYPE LOCALITY IN SAN ANTONIO SANDSTONE
After Royo y Gómez

Miocene or Pliocene

PEÑATA fauna; Late Miocene or Early Pliocene.
?San Antonio formation
(Stirton, 1946a)

Class—MAMMALIA
Order—RODENTIA
Family—Dinomyidae; *?Gyriabrus royoi* Stirton, 1946; holotype; upper premolar and incisor

The only representation of this fauna is an upper premolar and an incisor of a rodent found by Royo y Gómez in a light greenish-gray, fine-grained sandstone near kilometer 35 carretera de Toluviejo, near Sincelejo, Bolivar (Fig. 11). This land-laid formation is made up primarily of sandstones but also with considerable volcanic ash. Unfortunately the writer has not seen the section in this area, and after much searching in the literature and discussion

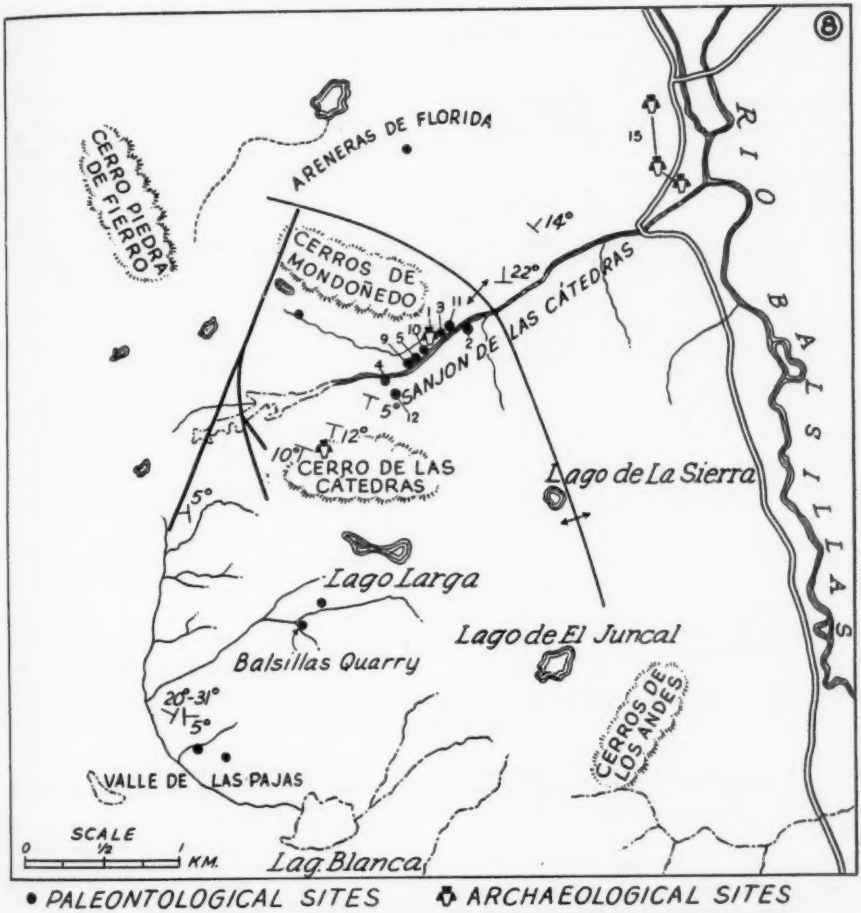

FIGURE 12.—MONDONEDO REGION, CUNDINAMARCA, SHOWING PALEONTOLOGICAL AND ARCHAEOLOGICAL SITES

See sections in Figure 13.

with others he has not been able to determine which of the names applied is referable to this unit.

It would seem to fall within the "San Antonio" sandstone of Beck (1921) and possibly in the "Savana" sandstone of Werenfels (1926). On the other hand the name "Sincelejo" sandstone was used by Werenfels for 200 meters of poorly consolidated yellow and gray conglomerate and sandstone overlying unconformably his "Savana." The writer doubts that the rodent came from this unit.

Below this continental formation are marine beds with mollusks and foraminifers said to be middle Miocene in age, though the lower part of the formation or group, as the case may be, is probably older. These marine beds apparently belong in Beck's (1921) "Huertas" group (called series by him), in Anderson's (1926; 1929) "Tubará" group, and in Werenfels' (1926) "Cerrito" formation.

CORRELATION: Fortunately the rodent premolar fits into the evolutionary series of the family Dinomyidae. It is clearly more advanced than the *Scleromys* and *Olenopsis* species in the La Venta fauna yet has not evolved into the giant rodents found in the Entre Ríos beds of Mesopotamian age in Argentina. It may be Chasicoan, or it may be a small species of Mesopotamian age nearer the equator. The "Gigante" tuffs overlying the beds containing the La Venta fauna and exposed between Villevieja and Neiva, Huila, may be of the same age.

The Sincelejo locality is of unusual interest because it gives us our first information on a marine and continental tie-in for northern South America

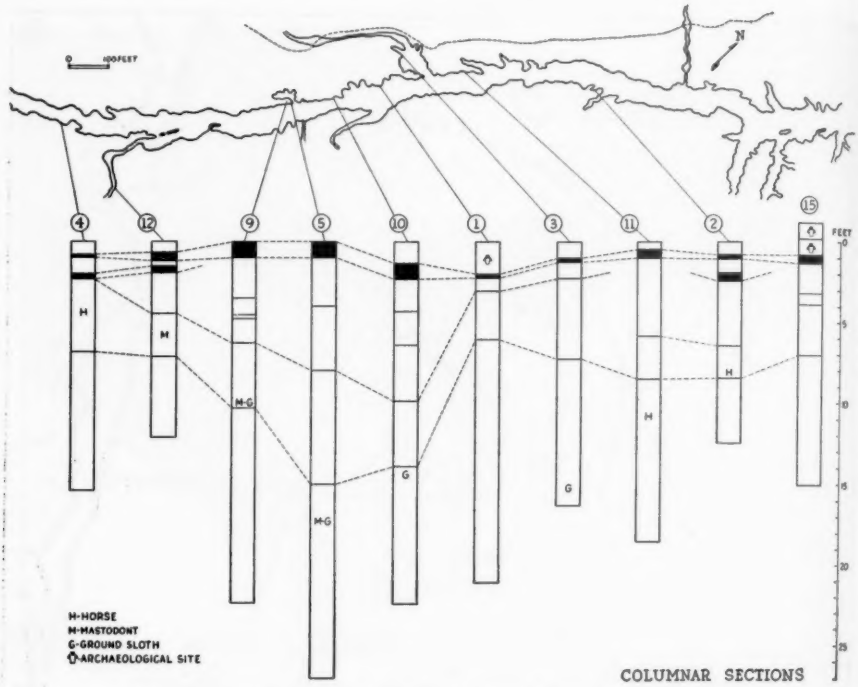

FIGURE 13.—SANJON DE LAS CATEDRAS, VALLE MEDIA, CUNDINAMARCA
See map of region, Figure 12.

with the possibility of a trans Caribbean-Gulf Coast correlation of continental beds. An effort should be made to obtain more fossils from the area.

?Pliocene

COCHA VERDE fauna; ?Late Pliocene. ?formation (Royo y Gómez, 1942a; Stirton, 1946a)

Class—MAMMALIA
Order—ARTIODACTYLA
Family—Tayassuidae. *Selenogonus narinoensis* Stirton, 1946a; holotype; posterior part of mandible with $M_{\overline{3}}$

This fragment of a peccary mandible was found by Royo y Gómez in a dark-green fine-grained sandstone near Cocha Verde, on the Tuquerres-Tangua carretera, Nariño. From its appearance it could be either late Pliocene or early Pleistocene. Perhaps other fossils could be found in this area to throw some light on the age of the thick volcanic series in the region.

Pleistocene

CATEDRAS fauna: Late Pleistocene. Mondoñedo formation

(Botero, 1936; 1937; Cuervo Marquez, 1938)
Class—MAMMALIA
Order—EDENTATA
Family—Mylodontidae. *?Mylodon*; various skeletal parts
Order—PROBOSCIDIA
Family—Gomphotheriidae. *Haplomastodon chimborazi* (Proano), 1922; mandibles, cheek teeth, upper tusks, and limb bones
Order—PERISSODACTYLA
Family—Equidae. *Equus*; part of cranium, mandibles and isolated teeth

The Cátedras fauna occurs in exposures in the Mondoñedo and Balsillas hills and grazing land about 20 kilometers northwest of Bogotá, Cundinamarca (Figs. 12, 13). In a general survey of the paleontology of Colombia, Gerardo Botero (1937) lists the mammals including those that were located in these hills, but it is difficult to know if all his fossils recorded from near Mosquera actually came from the hills to the southwest. Cuervo Marquez (1938) figures only the kinds found by us in 1951, though he mentions four kinds of mastodonts, three kinds of llamas, and a glyptodont. Some of these older collections are in the Colegio de Salesianos in Mosquera and in the Museo de La Salle in Bogotá.

The best localities are about 3–4 kilometers southwest of the village of Mosquera in Sanjón de las Cátedras and on the Santamaria property where the old Balsillas quarry can still be seen. These vertebrate-bearing beds are here designated as the **Mondoñedo formation** with the type locality in

Sanjón de las Cátedras, and the type of our Cátedras fauna is from the same beds.

The Mondoñedo formation in the hills and the Sabana formation on the Bogotá plain in part were probably laid down at the same time, but are distinct stratigraphic and cartographic units. The Mondoñedo beds were deposited in the washes, over the hill slopes, and in the land-locked basins or lakes in the hills. The thinning out of the beds from the valleys and basins up the hill slopes and other evidence indicate quite clearly that these sediments accumulated after the hills were uplifted. On the other hand the Sabana formation is composed of outwash from the mountains around the Bogotá plain, stream deposits, and laminated lake beds.

The Mondoñedo formation rests with angular unconformity on Cretaceous sandstones, claystones, and limestones from which the Pleistocene sediments have been derived. The thickest section measured in the Sanjón de las Cátedras was 6 meters. The color and the composition of the sediments depend on the kinds of Cretaceous sediments on the adjacent slopes, but they are usually sandy silts. The lower member contains angular chert and large sandstone boulders. The formation is capped with five or six buried soil profiles seldom more than 3 meters thick. These include one to three layers of black softer humic members and three thicker units of hardpan which are humic silty clay loams cemented with calcium carbonate. In the upper hardpan there is evidence of human occupation, but it has not been dated. Farther down the valley another cultural level rests on the upper hardpan. According to Luis Duque Gómez, Director del Instituto Etnológico in Bogotá, this is of conquest age. The Instituto Etnológico is still conducting work in these sites.

CORRELATION: The stratigraphic relationship of the fossil-bearing beds to the cultural levels, particularly the upper hardpan where a figurine and a stone ax were found, leads to the belief that the Cátedras fauna and the Mondoñedo formation are late Pleistocene. Until the sequence of Pleistocene vertebrates is better known in South America, fossils like those found will be of little help in detailed correlation.

Numerous Pleistocene finds have been made in different parts of Colombia, but for the most part they are isolated teeth or bones, and their stratigraphic significance does not fall within the scope of this report.

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GROUND-WATER BEHAVIOR IN THE HERSHEY VALLEY, PENNSYLVANIA

BY RICHARD M. FOOSE

ABSTRACT

The Hershey Valley, part of the Great Valley, is 9 miles east of Harrisburg, Pennsylvania. Beekmantown limestone (Ordovician), striking N. 70° E., and steeply overturned to the southeast, underlies the valley.

The Annville Stone Company mines the high-calcium Annville limestone (top of the Beekmantown) 1½ miles northeast of the Hershey Chocolate Corporation plant. For many years 3500 gpm was the normal pumping discharge. Neither industry adversely affected the other nor any of the valley residents.

Plans of the Stone Company to mine at lower levels prompted geologic studies by the Chocolate Corporation. One hundred wells and springs in the valley were systematically observed, 12 of the most critical being measured every 2 weeks. For each there was determined the water elevation; water temperature; pH value; ppm of dissolved solids; ppm of the Ca, Mg, PO_4, SiO_2, Al_2O_3, SO_4, and Fe ions; and yield in gpm of the springs. Details of ground-water movement were studied by using fluorescein. Water table maps of the valley were periodically made.

When the mine began pumping 6500 gpm from lower levels in May 1949, it drastically lowered the ground-water level over an area of 10 square miles, drying up most valley springs and many wells. About 100 sink holes of all sizes developed in the area, endangering life and property. The water-table map of November 1949 showed a huge cone of depression with the apex at the mine.

Ground-water levels were successfully restored over a large area when the Chocolate Corporation recharged water to the ground, beginning December 1949.

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Introduction

The Hershey Valley is a small part of the Great Valley, about 9 miles east of Harrisburg and 7 miles west of Lebanon, between Hummelstown and Palmyra, Dauphin and Lebanon counties, Pennsylvania (Fig. 1). Its north and south boundaries are low irregular hills that stand about 200 feet above the nearly flat and very gently rolling valley floor. The Hershey Valley comprises about 40 square miles, nearly all of which is highly productive farmland. Hershey, the main town in the valley, is well known as the home of the Hershey Chocolate Corporation, the world's largest chocolate plant.

Swatara Creek, the major stream in the valley, flows southwestward to enter the Susquehanna River 9 miles away at Middletown. The major tributary to the Swatara in this area is Spring Creek, a small stream with little variation in flow.

Acknowledgments

It is a pleasure to acknowledge the keen interest, constant aid, helpful suggestions, and pertinent criticism of the engineers and officials of the Hershey Chocolate Corporation during the field work and in the preparation of this paper. I am indebted to the Corporation for permission to publish these data. My colleague, George K. Biemesderfer, has participated in nearly all phases of the field and laboratory investigations since their beginning in 1947. Inevitably he has contributed a great deal to these studies; I am grateful to him not only for his geologic aid but also for his pleasant personality which made so much of the work agreeable. Donald U. Wise and Alan Geyer were of great help in preparation of the final drawings. Mr. Alan Zook, of Hershey, Pennsylvania, prepared the photographs.

I appreciate the sound and constructive criticism of V. T. Stringfield and Jack Graham who kindly reviewed the manuscript.

Stratigraphy

Ordovician

Beekmantown limestone.—The floor of the Hershey Valley is underlain by the Beekmantown limestone. Rocks of this formation vary widely in physical and chemical characteristics, but generally are dense, laminated, blue-gray magnesian limestones. Locally they are crystalline white limestones or dark-blue dolomites. More-detailed stratigraphic mapping could probably outline several distinct lithologic units within the Beekmantown.[1] One such unit, the Annville limestone, has already been mapped along the north side of the valley about 500–600 feet below the stratigraphic top of the Beekmantown. The Annville limestone is 250–300 feet thick at most places in the area and bears more than 95 per cent $CaCO_3$.

These valley limestones with their varying chemical characteristics have been differentially weathered and dissolved by percolating ground-water. Generally, the purer limestones have many more solution openings than the dolomites. Cavities and openings have been developed chiefly along the bedding and cross joints in the limestones and are naturally larger and more numerous near the surface (Fig. 2A). A 10-foot zone along the contact of the hanging wall of the overturned Annville limestone and the overlying dolomite is exceedingly permeable. At many places these openings are tightly plugged with mud, most of it residual soil, from the weathering of the limestones, that had fil-

[1] Recent work by C. E. Prouty (1951) and C. Gray (1952) break the Beekmantown into several discrete formations, not considered in detail here.

FIGURE 1. BASE MAP OF THE HERSHEY VALLEY

tered down into the openings and become packed.

Martinsburg shale.—Conformably overlying the Beekmantown limestone, the Martinsburg shale is tan and very thin-bedded with occasional thin beds of sandstone, limestone, and chert. It weathers to characteristic light-brown or greenish-brown hackly fragments. The Martinsburg shale underlies the northern half of the Great Valley and makes the irregular hills that form the northern boundary of the Hershey Valley.

Triassic

Gettysburg shale.—Soft red shales and poorly cemented silty sandstones with a few interbedded conglomerates bearing large white and pink quartz pebbles constitute the Gettysburg shale which underlies the hills bounding the Hershey Valley to the south.

Structure

The Beekmantown limestone and Martinsburg shale strike generally N. 65°–70° E. and are steeply overturned to the south, having average dips of 45°-65° S. The limestone especially has yielded to severe deforming stresses so that there are many tight folds, some isoclinal, some recumbent, and, it is suspected, nappelike structures are locally developed. Boudinage (sausage-shaped remnants of beds) is beautifully developed throughout the formation, attesting to the great amount of stretching and flowage.

A prominent set of cross joints and a well-developed set of longitudinal joints are developed in the limestone and to a much lesser extent in the shale. Solution of the limestone has been greatest along the cross joints and the steep south-dipping bedding, and to a lesser extent along the longitudinal joints.

Several faults displace the limestone and shale. Some of these strike northeast and have displaced beds as much as 3000 feet at the surface. Others are cross faults, striking generally north-northwest. Several of the latter cut into the mining operations of the Annville Stone Company and the H. A. Millard Company. Locally, along these faults, the rocks have been extensively brecciated. A connecting series of steeply dipping normal faults separate the Beekmantown limestone along the south side of the Hershey Valley from the younger Triassic rocks to the south. The Triassic rocks have relatively uniform dips of 25°–35° NW.

Mining

The high-calcium Annville limestone has been quarried and mined for many years at many places along the Great Valley. In the immediate area it is being quarried by The H. A. Millard Co. at Swatara Station a mile southwest of Hershey. Their quarry is about 125 feet deep. One and a half miles northeast of Hershey, the Annville Stone Company[2] operates a mine whose workings extend more than 400 feet below the surface. The rock is removed by a combination of room and pillar and shrinkage stope mining methods. The main slope into the mine has its origin on the floor of a large quarry, about 125 feet deep, which was previously operated by the same company.

Ground-Water Geology

Springs

In the vicnity of Hershey, and in the valley of Spring Creek, the major tributary to Swatara Creek, there were once many springs, several of them yielding thousands of gallons per minute. One large spring in particular made several large ponds a few hundred feet east of the plant of the Hershey Chocolate Corporation. One reason for the plant's location was this excellent source of water, known first as the Derry Spring. For more than a century the various springs were used on many of the farms in the valley as their only water supply, and for nearly 50 years the Derry Spring was used by the Hershey Chocolate Corporation. In recent years four of the largest springs, located about a mile southeast of Hershey, have supplied water for livestock and for the domestic needs of large numbers of orphan boys living on farms maintained by the Hershey Estates. As far as it has been possible to determine, none of the large

[2] This company was purchased by the Michigan Limestone and Chemical Company in November 1951. In December 1952 it is operated as the Michigan Limestone Division of the United States Steel Corporation.

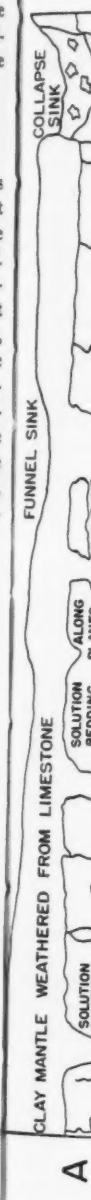

FIGURE 2. IDEALIZED DIAGRAMS OF LIMESTONE AND GROUND-WATER CONDITIONS

springs ever went dry, even during the longest droughts.

Almost all the springs emanate from solution-widened bedding or from the intersection of a prominent cross joint with a bedding plane. The large Derry Spring is situated on the Annville limestone, which has more and larger solution openings than the rest of the Beekmantown limestone. The other large springs in the Hershey Valley are either along the outcrop belt of the Annville limestone or near it and connected with it by well-developed cross joints.

Wells

During the past 50 years many wells were drilled on farms and on properties where new homes were built. Except for very shallow wells, in which the water never stood more than several feet above the bottom of the well, none of these are known to have gone dry. An exhaustive study of nearly 200 such wells located throughout the entire valley would indicate that lowering of the water table due to drought was never greater than 10 feet. A series of 12 wells were also drilled by the expanding plant of the Hershey Chocolate Corporation between 1926 and 1948. Yields ranged from 200 to 1200 gallons per minute, and a few had no yield at all, demonstrating how poor an aquifer limestone can be "in between the fractures." The Corporation wells continued to have a constant yield in wet and dry season alike, and they apparently never affected in any way the adjacent springs or wells in the valley.

The drilled wells in the valley range from 50 to 400 feet in depth, most of them averaging about 150 feet. Although nearly all of the 190 wells studied had a yield adequate for their demands, about a third had apparent yields of only 2–5 gallons per minute, and about a dozen had essentially no yield. Average yield is 10–30 gallons per minute. Most of the wells with larger yields are along the outcrop zone of the Annville limestone, in or near fractured rock associated with the faults, or relatively near either of these two water-bearing zones and connected to them by joints. This further reveals that many more and larger solution openings were developed in the purer limestone. A few such wells are located in other high-calcium beds of the Beekmantown limestone in the middle and southern parts of the valley.

Several wells that initially had a small yield produced very muddy water when pumped hard and long. These wells ultimately produced clear water and with much greater yield. Obviously the well had intersected mud-filled solution openings along either bedding or joints. Continuous pumping had eventually flushed out the mud and opened up to easier access the many connecting passages through which ground water could move toward the well. In several such instances, the water level in an adjacent well or wells was lowered, indicating that the mud seal had maintained a perched or elevated water table in the local area (Fig. 2C). In several wells the ground-water level stands many feet above the regional water table; this is attributed to isolation by mud seals or by unusually tight fractures.

Quarries

The Millard quarry a mile southwest of Hershey pumps an average of 450–500 gallons per minute to maintain their operation under normal conditions.

The Annville Stone Company mine 1½ miles northeast of Hershey pumped 3000–3500 gallons per minute (until 1949) to maintain their operation under normal conditions. After that company grouted the openings adjacent to the mine in the spring of 1950, the average pumping rate was 2500 gallons per minute[3].

Normal Water Table

Examination of the mass of data compiled for the wells and springs of the valley revealed that the water table bore a normal relationship to the topography, the surface drainage, and the rock types. It stood considerably higher in the Martinsburg shale to the north and in the Triassic rocks to the south than it did in the limestone of the valley. It sloped gently toward the west in the direction of Swatara Creek, the major surface stream.

[3] During January to March 1952 the average pumping rate was again about 3200 gpm. Following further grouting during the summer, the rate was 2500 gpm. in December 1952.

The effect of the many individual wells on the water table was entirely local throughout the valley. Each well had its own cone of depression of varying size depending on the amount of water pumped, the depth at which the pump intake was established, and the potential yield of the well.

The effect of the mine and the quarry operations and of the cluster of wells at the Hershey Chocolate Corporation was greater. The cones of depression surrounding these three places had considerable area and depth. The smallest of the three was that of the Hershey Chocolate Corporation. That cone of depression was the result of the combined cones of nine wells; its area was about 50 acres, and the total depth of the cone was about 15 feet. The cone surrounding the Annville Stone Company mine was elongate, covered an area of at least 255 acres, and was more than 50 feet deep. The cone surrounding the Millard quarry was somewhat more localized, covering about 165 acres, and was about 40 feet deep. In each case the depth of the cone is the difference between the lowest measured water level and those water levels not affected by pumping. The boundary of the cones was likewise determined by water-level measurements that showed no change due to pumping.

Some wells even within these well-established cones of depression were apparently completely dissociated from the others; their static water level was generally higher than the surrounding water table, and some could be pumped down lower than the surrounding water table with no apparent widespread effect. All this indicates the degree of "tightness" of the limestone or the effectiveness of mud seals in the fractures, or both.

When the first water-table map of the Hershey Valley was prepared in October 1948 (Fig. 3) the conditions throughout the valley were essentially the same as they had been for many years. This was true of both mine and quarry operations which had for many years pumped the same amount of water, and also of the Hershey Chocolate Corporation. In addition, it had been neither unduly dry nor wet during the summer and fall months. The 1948 water table could therefore serve as a frame of reference for future observations.

Ground-Water Movement

Normal relationship.—The normal movement of ground water throughout the valley prior to any wells being drilled or quarries opened must have been toward the southwest in the direction of the gently sloping groundwater table and into the valleys of Swatara Creek, Spring Creek, and their tributaries. As of 1948, the direction of ground-water movement was locally controlled by the many wells and by the larger industrial operations. However, no single well or group of wells nor the quarry and mining operations adversely affected the supply of any other source of water at that time.

Annville Stone Company blast of August 1946.—During routine mining operations in August 1946, a blast in the hanging wall of the mine exposed a 6-inch wide solution channel about 375 feet below the surface out of which poured an estimated 8000–10,000 gallons of water per minute, flooding the mine in the course of a day. When this occurred, near-by wells dried up, ground-water seepage into an adjacent quarry ceased, the Derry Spring 1½ miles southwest dried up on the second day, and the water level in two wells of the Hershey Chocolate Corporation rapidly declined. After many months of labor the opening in the mine was sealed off with a large steel plate, the adjacent wells had water in them again, the flow of the Derry Spring was restored, and water levels in the Corporation wells were again normal. This incident effectively showed that there were solution openings in the limestone connecting the Derry Spring and some of the Corporation wells with the mining operation.

Annville Stone Company pumping test of August 1948.—From August 30 to September 4, 1948, the Annville Stone Company pumped an average of 5500 gpm, holding the water at a level of about 345 feet below their quarry floor[4] as a test preliminary to permanent installation of pumps for deeper mining operations. On September 2 the newly drilled Derry Spring well (drilled at the side of the Derry Spring. Yield: 2100 gpm) had "dried up"; the water level fell from an elevation of 355 feet to 313

[4] Personal communication from Mr. Lyles, mine superintendent, in September 1948.

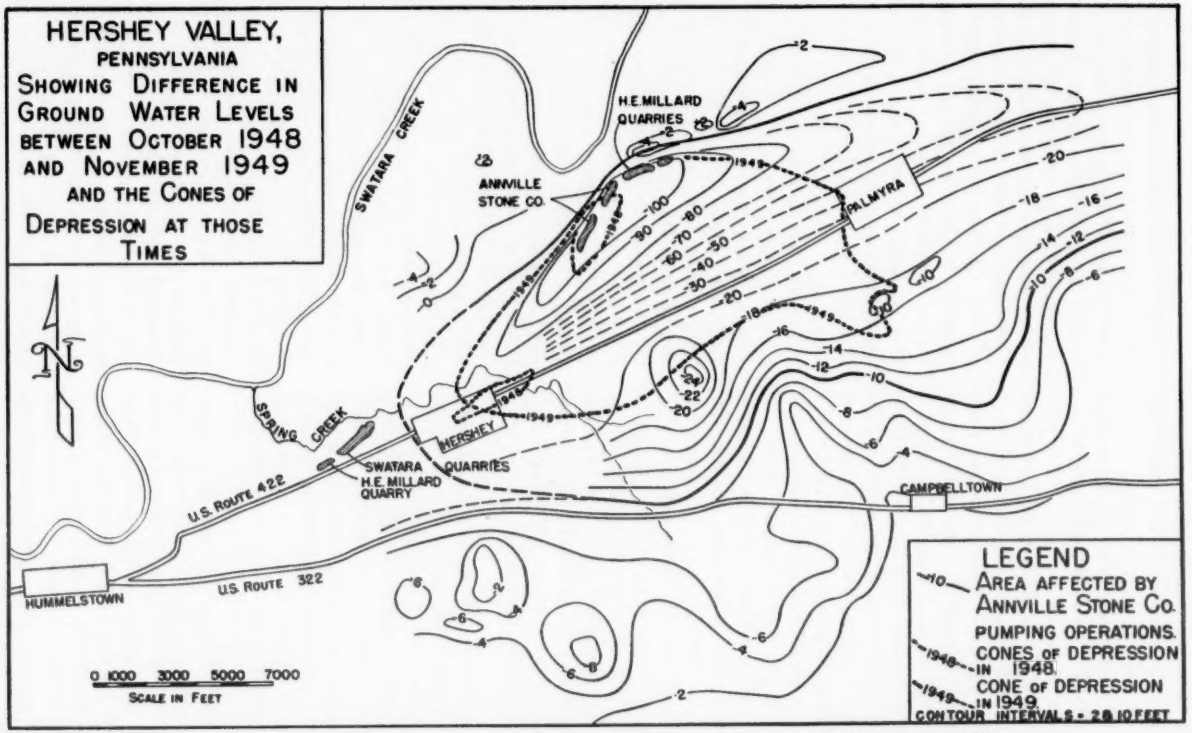

FIGURE 3. GROUND-WATER LEVELS IN THE HERSHEY VALLEY, OCTOBER 1948–NOVEMBER 1949

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feet, which was below the pump intake. In another Corporation well, the water level was lowered more than 30 feet, and it too was "dry." The connection between the wells and the mine was obvious. On September 8 the water level began to rise in the two wells. Within a couple of days normal pumping operations could be resumed.

After the 1946 incident and the August 1948 pumping test, it seemed obvious that there was easy movement of ground water through the many cavernous passageways of the Annville limestone, especially in the 10-foot wide zone at the hanging-wall contact with higher magnesian rock. Within 4 days the normal level of ground water was drastically altered 1½ miles from the mine.

Ground-Water Fluctuations

Geologic studies begun in 1947 were intensified after the pumping test. It was obvious that permanent installation of the pumps at a lower level in the mine would endanger the water supply not only of the Chocolate Corporation but of farms and residents throughout a large portion of the valley. The Hershey Chocolate Corporation inaugurated a program designed to learn everything possible about the behavior of ground water in the entire valley. This was more than a study for augmenting or protecting its own water supply; it was conducted in the interest of all the residents of the valley and with the ultimate goal of conserving vitally needed grounded water.

During the fall of 1948 an exhaustive survey was made of practically every well and spring in the Hershey Valley. For about 200 wells and 30 springs, complete information and a detailed "personal history" were compiled. Answers to the following list of questions were sought from the well owners:

1. Location of residence (on base map of valley).
2. Well present?
3. Do neighbors have wells?
4. Are wells dug or drilled?
5. Name and address of well driller?
6. Depth of well in feet?
7. Diameter of well?
8. Is well cased?
9. Depth of casing in feet?
10. Normal water level (feet below collar)?
11. Yield in gpm?
12. Depth of intake?
13. Has level fluctuated?
14. Was well ever dry?
15. Was there a drought when dry?
16. Was the water from well ever muddy?
17. Was water from well ever discolored?
18. Did water ever taste strange?
19. Any unusual behavior?
20. Is the well accessible?
21. Consent to observe and measure well (yes or no)?

From these were selected about 100 wells and springs which, because of their characteristics and their location, were ideal observation points (Fig. 1). Elevation of the wells and springs was surveyed, and each was carefully measured to determine: (1) depth to water level; (2) elevation of the water table; and (3) temperature of the water.

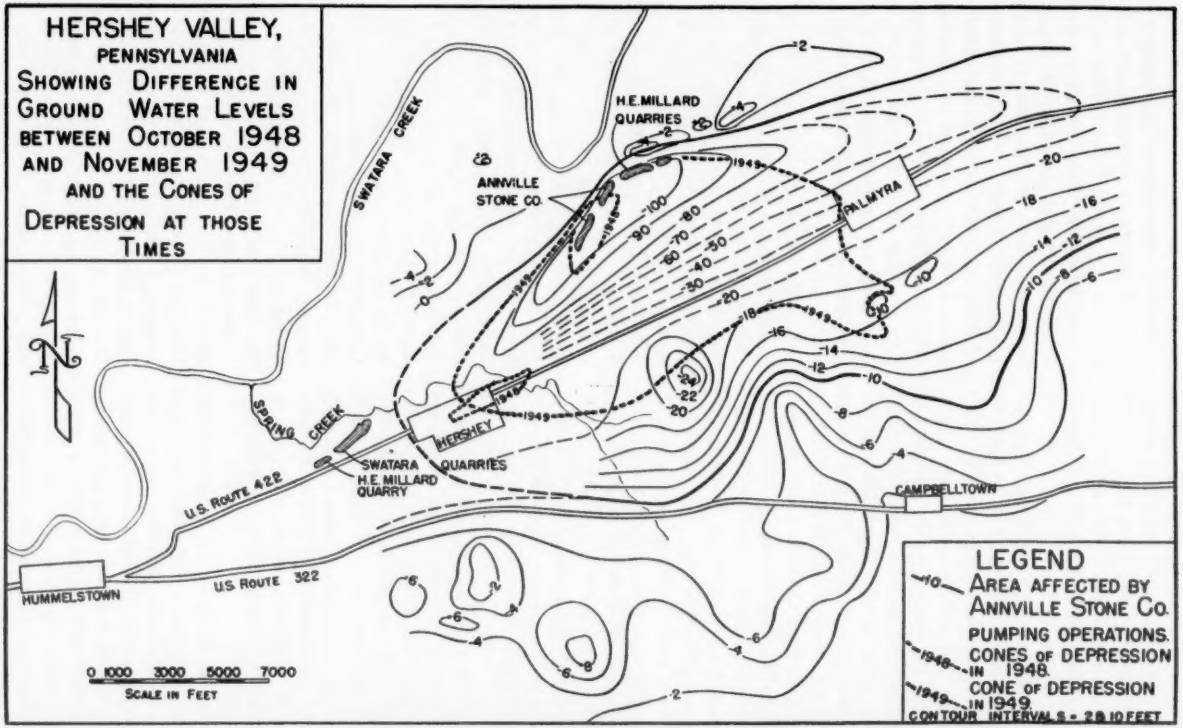
Water samples were collected in pint-sized, waxed, cardboard containers and sealed for the following laboratory tests: (1) pH value. Hydrogen-ion concentration; (2) total parts per million of dissolved solids; and (3) parts per million of the following chemical ions in the water: Ca, Mg, PO_4, SiO_2, Al_2O_3, SO_4, and Fe. All tests were made within 48 hours and as soon as possible. For the springs, approximate or accurate yield in gallons per minute was determined.

The measurements were made at all observation points in October 1948, November 1949, May 1950, November 1950, May 1951, November 1951, May 1952, and November 1952.

The results of water-table measurements made during October 1948 are summarized on Figure 3. These measurements were made when all conditions were similar to those that existed throughout the valley prior to August 1946 and might be termed "normal conditions."

During May 1949 the mining operation inaugurated its new pumping program designed to permit deep mining at substantially lower levels. About 6500 gallons per minute became the normal discharge from pumps whose intake was more than 330 feet below the surface. The results of this pumping program were phenomenal and critical. Near-by wells were either dried

FIGURE 3. GROUND-WATER LEVELS IN THE HERSHEY VALLEY, OCTOBER 1948–NOVEMBER 1949



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21. Consent to observe and measure well (yes or no)?

From these were selected about 100 wells and springs which, because of their characteristics and their location, were ideal observation points (Fig. 1). Elevation of the wells and springs was surveyed, and each was carefully measured to determine: (1) depth to water level; (2) elevation of the water table; and (3) temperature of the water.

Water samples were collected in pint-sized, waxed, cardboard containers and sealed for the following laboratory tests: (1) pH value. Hydrogen-ion concentration; (2) total parts per million of dissolved solids; and (3) parts per million of the following chemical ions in the water: Ca, Mg, PO_4, SiO_2, Al_2O_3, SO_4, and Fe. All tests were made within 48 hours and as soon as possible. For the springs, approximate or accurate yield in gallons per minute was determined.

The measurements were made at all observation points in October 1948, November 1949, May 1950, November 1950, May 1951, November 1951, May 1952, and November 1952.

The results of water-table measurements made during October 1948 are summarized on Figure 3. These measurements were made when all conditions were similar to those that existed throughout the valley prior to August 1946 and might be termed "normal conditions."

During May 1949 the mining operation inaugurated its new pumping program designed to permit deep mining at substantially lower levels. About 6500 gallons per minute became the normal discharge from pumps whose intake was more than 330 feet below the surface. The results of this pumping program were phenomenal and critical. Near-by wells were either dried

FIGURE 3. GROUND-WATER LEVELS IN THE HERSHEY VALLEY, OCTOBER 1948–NOVEMBER 1949

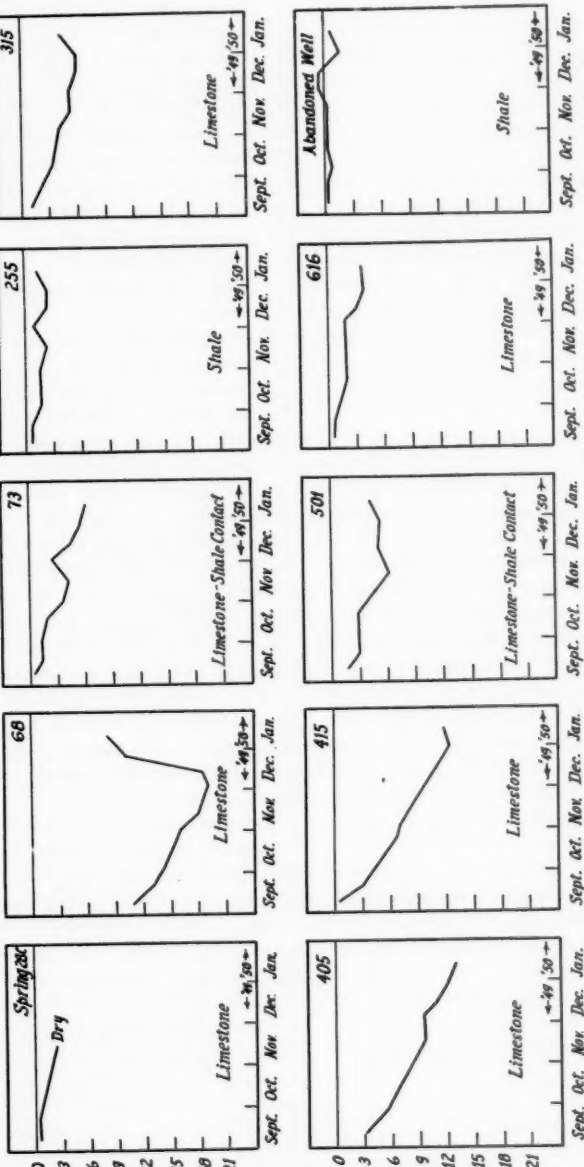


FIGURE 4. ELEVATION CHANGE OF WATER TABLE IN FEET IN 10 WELLS AND SPRINGS OBSERVED BIWEEKLY DURING 1949–1950
(Referred to October 1948 levels)

up or their water levels rapidly declined. The Derry Spring well was effectively "dried up." Water levels in two of the Corporation wells immediately declined, and, with time, others were similarly affected. The large springs on the farms dried up. Spring Creek dried up. Irrigation water was not available, and crops suffered. Throughout the valley, water levels in wells dropped at an alarming rate, and many wells went dry. Families had to haul water.

Biweekly well-observation program.—To maintain a more constant check on all conditions, a selected group of nine wells and three springs (Fig. 1) were measured every 2 weeks between September 1949 and the spring of 1951, and the water-surface elevations were referred to October 1948 levels. These wells and springs were selected on the basis of their geographic position with respect to the Annville Stone Company and the Hershey Chocolate Corporation and their geologic location with respect to the Martinsburg shale, the Annville and Beekmantown limestones, and the Triassic rocks.

The rate of decline of water levels in the wells and springs during the fall of 1949 far exceeded the rate or amount of decline that might be expected from the lower than average rainfall the area received at that time (Fig. 4). The average lowering of the water table caused by lesser rainfall is shown by those wells located in the Martinsburg shale, in the Triassic sandstone, and in those parts of the Hershey Valley so situated that the wells could not be affected by the pumping operations of the Annville Stone Company. Typical wells of the latter type are Nos. 255, 616, and the abandoned well. Other wells which are located geographically where they could be partly affected by pumping operations of the Annville Stone Company, such as Nos. 73, 315, and 501, suffered greater decline of the water level. Wells that were badly affected by the pumping operations of the Annville Stone Company because of closer proximity and underground channels connecting them with the operations are Nos. 68, 405, and 415 and springs 28A and 28D.

The effect of ground-water recharge during December 1949 and January 1950 is also shown on Figure 4. The beginning of the restoration of the ground-water table to normal levels by May 1950, following the grouting program of the Annville Stone Company, is indicated.

Decline of springs.—Another example of the relatively minor influence of rainfall on the rapidly declining levels of wells and springs is the combined flow of water from spring 28A and the Groy spring. Early in September 1949 a weir was erected at the mouth of a pond created by these two springs to measure the flow of water (Fig. 5). Between September 12 and 15 about 415 gallons per minute flowed through the weir. Between September 15 and October 24, the flow of water steadily declined to less than 25 gallons per minute. The rate of decline was fairly constant. Only after periods of relatively heavy rainfall did the rate of decline alter, and only after the very heavy rain of October 25 did the flow of water actually increase, and then only for one day, to approximately 40 gallons per minute. On November 2, no more water flowed over the weir, and 2 weeks later the floor of the spring pond was dry (Pl. 1, fig. 1). This phenomenal decline in flow can in no way be attributed to the low rainfall of the period because the springs had never dried up before, even during times of severe drought. Two other large springs, 28C and 28D, located near by dried before the end of November 1949.

Precipitation data.—Precipitation data were assembled for all years between 1930 and 1950 from Federal weather bureau stations at Harrisburg (9 miles west) and Lebanon (14 miles east) (Figs. 6, 7). They show that during the 20-year period there were 5 years that were drier at Harrisburg and 4 years that were drier at Lebanon; in 2 other years rainfall at both places was almost the same. During none of these years did a spring dry up nor did any well in the valley exhibit a drastic lowering of water level.

Development of sink holes.—During the second month of the mining operation's new pumping program, sink holes began to form in the valley of Spring Creek. At first they were filled, but they continued to form so rapidly it became impossible to keep pace. They ranged from about 1 foot to 20 feet in diameter and from 2–10 feet deep. All of them had very steep walls, as though the bottom had suddenly been pulled out and the unsupported earth above had quickly slumped (Pl. 1, fig. 2). Essentially

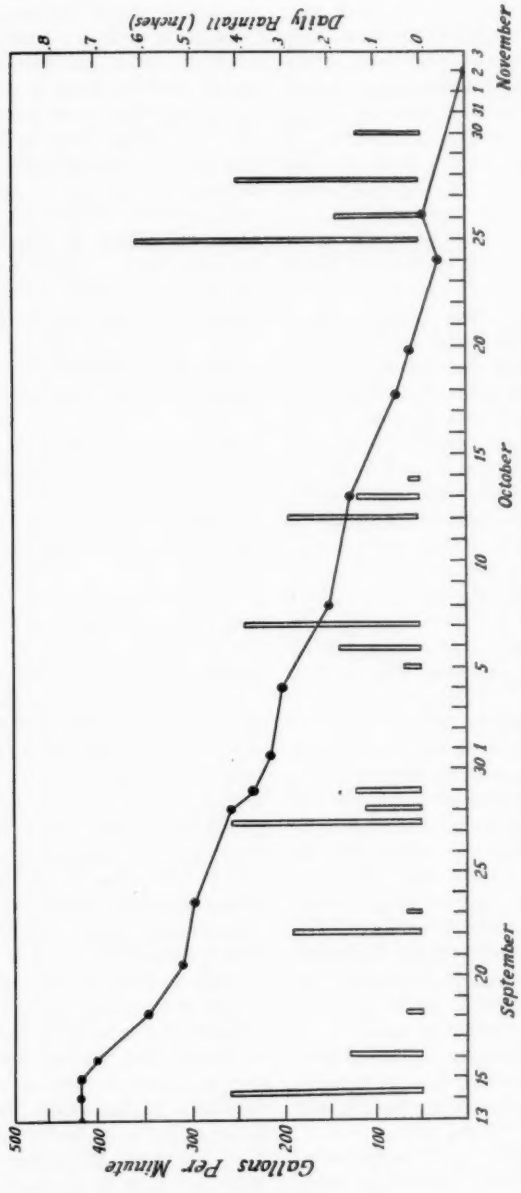


FIGURE 5. FLOW RECORD OF SPRING 28A DURING LATE 1949

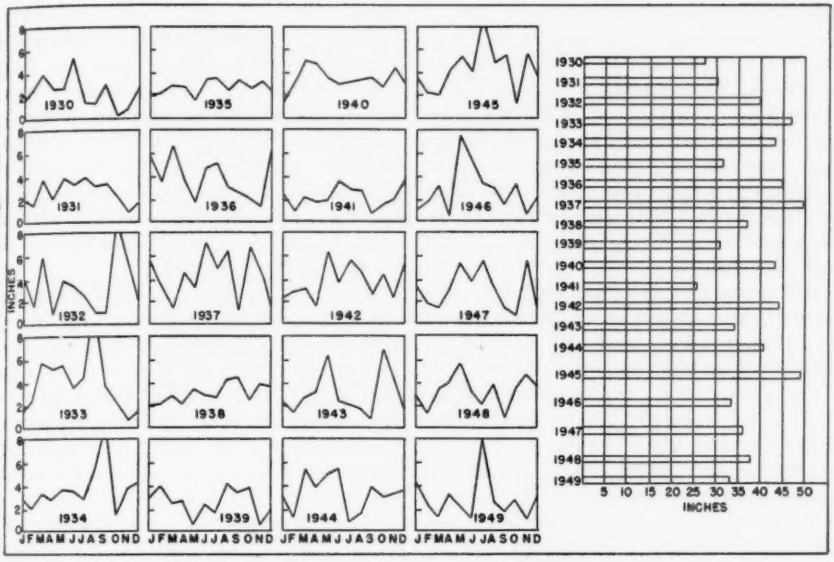


FIGURE 6. MONTHLY AND YEARLY PRECIPITATION AT HARRISBURG, PENNSYLVANIA
(Average yearly precipitation is 37.94 inches)

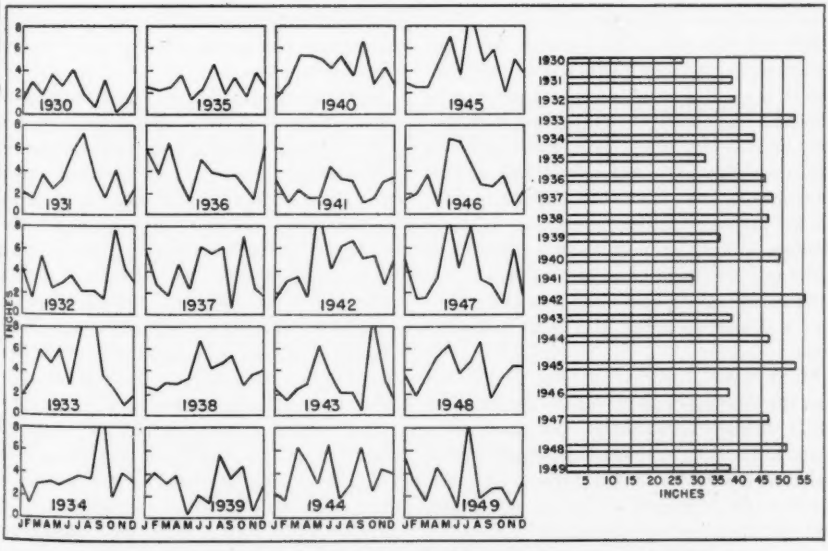


FIGURE 7. MONTHLY AND YEARLY PRECIPITATION AT LEBANON, PENNSYLVANIA
(Average yearly precipitation is 43.50 inches)

that is what happened. Instead of the ground-water table being near the level of Spring Creek, where it normally is, it had been lowered by many feet. Water flowing in the channel of Spring Creek moved downward through whatever openings were available, carrying along large quantities of silt, sand, and gravel. Unsupported areas in both the channel of the creek and adjacent to it, along its banks and on the flood plain of the creek, slumped and made sink holes (Fig. 2B).

Half a dozen of the nearly 100 sink holes formed in the bed of the creek. In one case an area 2½ feet in diameter on the surface of the silt under about 2 inches of water caved, leaving a pool 1½ feet deep. Around the edge of the pool the silt was warped down, showing concentric fractures. Water was obviously moving into the incipient sink hole. When probed with a stick it was apparent that no silt remained in the center of the hole. Only gravel was felt; all the silt had been carried downward. The next day this hole was 7 feet in diameter and 6 feet deep. Bedrock was exposed at the bottom of the hole showing a vertical fracture widened by solution. Into the hole poured all the remaining water of Spring Creek.

Prior to July 1949 very few sink holes had opened in the Hershey Valley within the memory of man. A careful survey of all the old existing sink holes, some dating back to before 1900, proved that more had opened up in the late summer of 1949 than existed before. Furthermore, whereas the older sink holes had formed at many places throughout the large Hershey Valley, some of them located along faults and prominent cross joints, the new ones had formed entirely within the area where there was a drastic lowering of the ground-water table.

Many valley residents attested to the fact that the older sinks all had gently sloping sides and a distinct funnel shape even when newly formed, whereas the 1949 sinks had precipitous walls and more of a cylindrical shape, indicating a very rapid lowering of the ground-water table and withdrawal of its support.

As a direct result of the large number of sink holes, the entire lower course of Spring Creek down to the plant of the Hershey Chocolate Corporation was dry after September 21, 1949 (Pl. 2). The existence of the sink holes and the rapidity with which they formed constituted a physical danger to life and property. Many areas had to be roped off in the valley of Spring Creek; the abutment of a highway bridge had to be reinforced when a sink hole developed under it; the possibility of serious property damage and failure of foundations was great. The hazard was not eliminated until after the successful raising of ground-water levels by recharging water to the ground.

A large spray pond is maintained just east of the Hershey Chocolate Corporation buildings in the bed of Spring Creek for the purpose of cooling industrial water prior to recirculation. During early November 1949, the level of water in the spray pond became very low because much of the water flowed in a reverse direction (upstream) along the dried-up channel of Spring Creek and sank into small sink holes in the channel. To prevent this a coffer dam was built at the eastern end of the pond on November 10. A constant danger was the imminent development of one or more sink holes in the bottom of the spray pond or underneath the corporation buildings. Although the sink holes developed in an increasingly larger area week by week, none reached the spray pond or the Corporation buildings.

Direction and speed of ground-water movement.—Coincident with the program of well and spring measurements, detailed studies were conducted to determine the physical characteristics of ground-water movement in the immediate vicinity of the Hershey Chocolate Corporation. The earliest tests, during the fall of 1948, were designed to show the nature of movement within the shallow cone of depression created by the Chocolate Corporation wells. Small charges of 1 and 2 pounds of fluorescein in alkaline solution were flushed into various wells; on arrival at other pumping wells observers noted: (1) time of arrival of dyed water; (2) duration of its observance; (3) intensity of coloration. Samples revealing the presence of fluorescein were collected at definite time intervals. Ultimately a fairly good picture was gained of the general direction of ground-water movement under various physical conditions, the speed of its movement, and something of the nature of the passages through which it

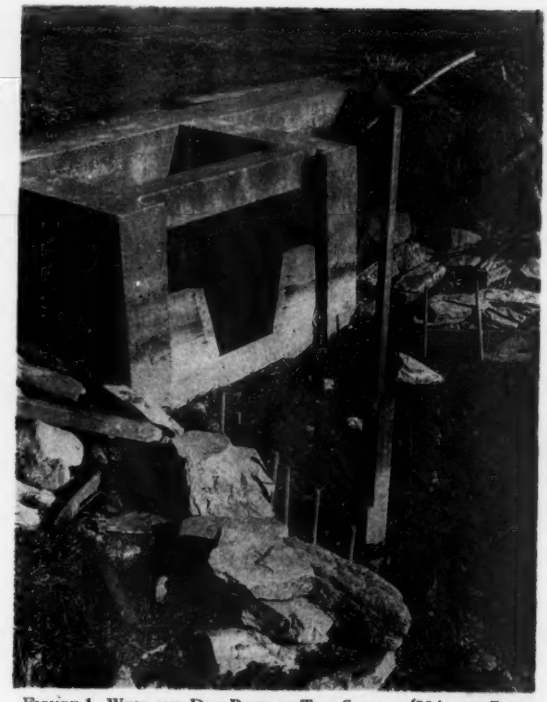


FIGURE 1. WEIR AND DRY POND OF TWO SPRINGS (28A AND GROY), NOVEMBER 1949

FIGURE 2. SINKHOLE IN FLOOD PLAIN ADJACENT TO SPRING CREEK

SINKHOLES AND DRIED SPRINGS

FIGURE 1. SPRING CREEK IN 1948

FIGURE 2. DRY BED OF SPRING CREEK, SEPTEMBER 1949

moved. These tests established the fact that most wells and springs are interconnected, though not always by channels through which there is easy movement. Within the Hershey cone of depression easiest movement is along cross joints and secondarily along bedding.

Beginning in July 1949 after the Annville Stone Company had begun its new pumping program a series of larger tests using dyed water were made to establish the direction and speed of ground-water movement throughout the valley. The results of those tests can be summarized:

(1) Ground water east of the Hershey Chocolate Corporation moved toward the Annville Stone Company mine, covering the full distance between 40 and 48 hours.

(2) Within an area extending no farther than 1000 feet east of the Hershey Chocolate Corporation, a small portion of the ground water moved toward the corporation wells that were being pumped, despite the fact that the water table had a regional gradient sloping toward the Annville Stone Company.

(3) Ground water moving along north-south cross joints toward Chocolate Corporation wells had the least interference from Annville Stone Company pumping operations. Greatest interference was in the general northeast-southwest direction along bedding.

(4) As ground-water levels were lowered throughout much of the Hershey Valley, many of the existing mud "seals" in cavities along joint planes and bedding were flushed out, thereby increasing the ease and speed of movement of ground water along them (Fig. 2C).

Water table in November 1949.—In November 1949 a new water-table map was drawn from measurements on all the wells and springs. The new ground-water table had a tremendous depressed area, which encompassed the Hershey Chocolate Corporation, a large part of the Hershey Valley, and centered about the mining operation (Fig. 3). Total depth of the depression was close to 200 feet. The pumps in the mine were set deeper than the bottom of the cone and were at all times pumping under head. Another map was drawn to show the total area affected by the new pumping program. In more than 10 square miles of the Hershey Valley the ground-water surface had been lowered more than 10 feet (Fig. 3). The investigation also proved that maximum lowering of the ground-water surface due to a lesser rainfall during the year was only 6–8 feet and that average lowering was 2–4 feet, these changes being observed in the areas underlain by Martinsburg shale and the Triassic rocks where no effect of the pumping operations was felt.

The relationship between the water-table levels of October 1948 and November 1949 is further shown by Figures 8 and 9.

Geochemical Tests

Hydrogen-ion concentration.—It was thought that a possible relationship between pH and ground-water movement might be revealed by mapping the values. Careful study of many such measurements indicates the following:

(1) Many measurements are necessary before any significant patterns emerge. With a large number, however, definite patterns are developed.

(2) An area of higher pH values follows the Annville limestone. This high "plateau" of hydrogen-ion concentration probably reflects the alkaline character of the underlying bedrock and possibly also the ease with which it is dissolved.

(3) The linear pattern of pH values tends to follow the established directions of easiest ground-water movement in the valley. This is not shown very well by Figure 10; however, it is indicated by comparative studies using maps prepared at different times.

(4) Differences in pH values rather sharply express the boundary between major rock types, such as the contact of the limestone with the Martinsburg shale and with the Triassic rocks. Figure 10 is a typical map showing average conditions. Changes in ground-water movement are revealed by accompanying changes on the map. Comparative studies are essential to an understanding of the changes.

Dissolved solids concentration.—A study of dissolved-solids concentrations gave the following results:

(1) It is necessary to have many measurements for much value to accrue from them.

(2) Where there are unusually high concentrations the ground water is "ponded" and

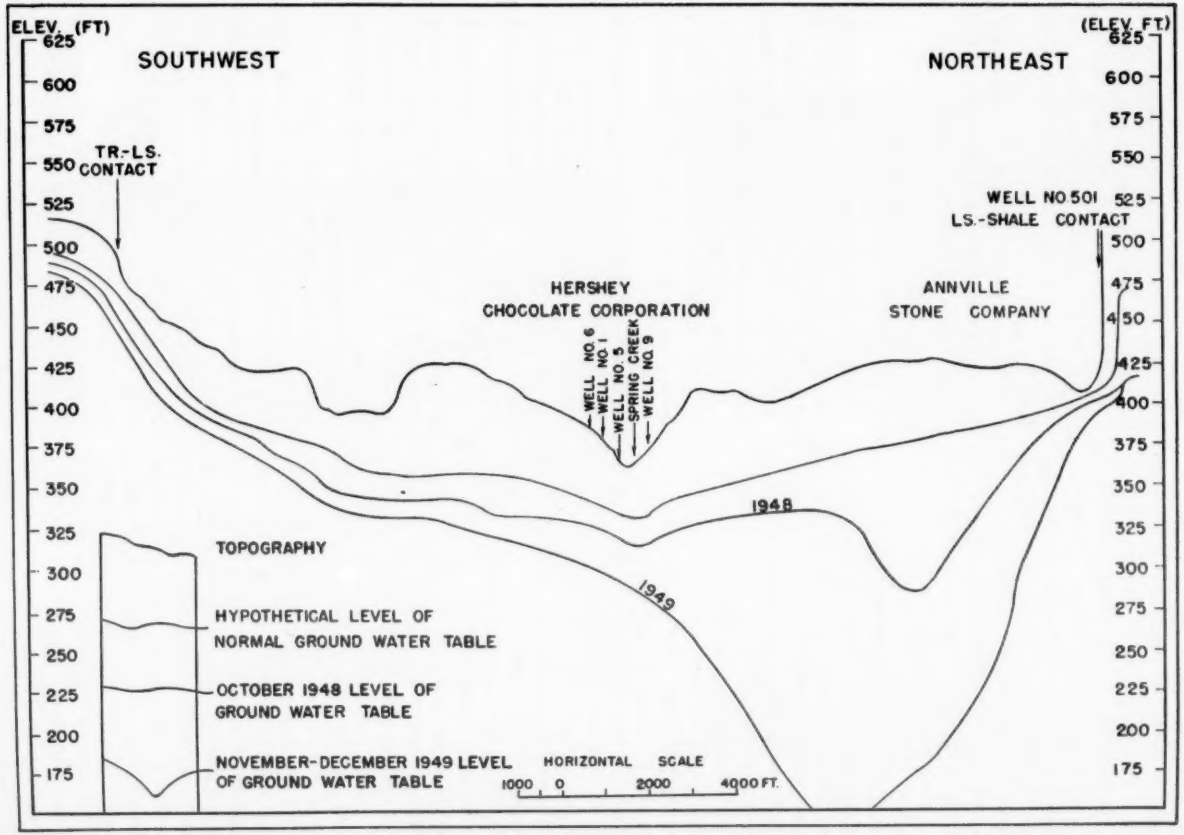
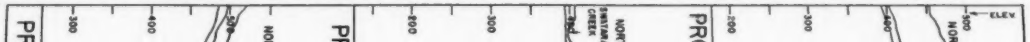


FIGURE 8. PROFILE D–D′ SHOWING RELATIONSHIP OF TOPOGRAPHY TO WATER-TABLE LEVELS IN 1948 AND 1949
ALONG THE SOUTH SIDE OF THE HERSHEY VALLEY



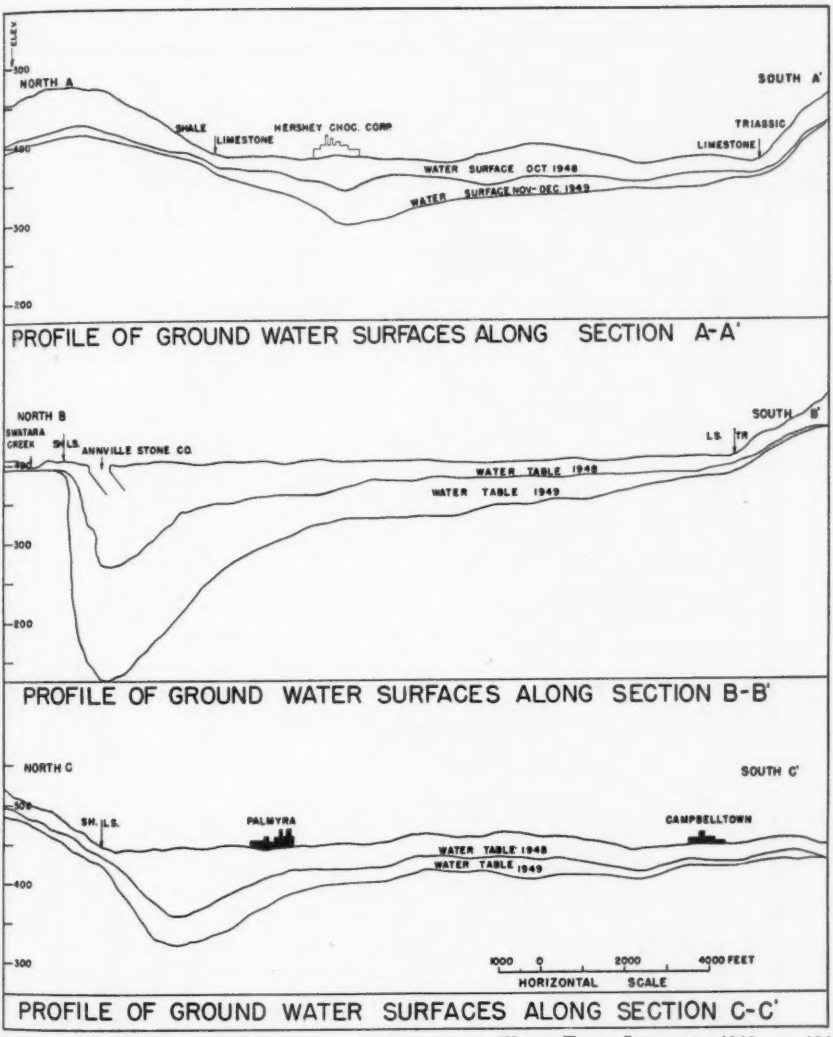


FIGURE 9. PROFILES SHOWING RELATION OF TOPOGRAPHY TO WATER-TABLE LEVELS IN 1948 AND 1949 TRANSVERSE TO THE STRIKE OF THE ROCKS AND THE HERSHEY VALLEY

FIGURE 8. PROFILE D–D' SHOWING RELATIONSHIP OF TOPOGRAPHY TO WATER-TABLE LEVELS IN 1948 AND 1949 ALONG ONE STRETCH OF THE HERSHEY VALLEY

moves very slowly or not at all. In such areas the water table may be perched and held back by mud "seals."

(3) Where there are uniformly low concentrations, the movement of ground water is easiest and the path of movement is generally defined by the low values. In the Hershey Valley this is along the Annville limestone, with very good movement along several faults and prominent cross joints.

(4) Differences in values of concentration reflect fairly well the boundary between the major rock types.

Figure 11 shows average conditions. Changes in ground-water movement are revealed by accompanying changes in the map. Comparative

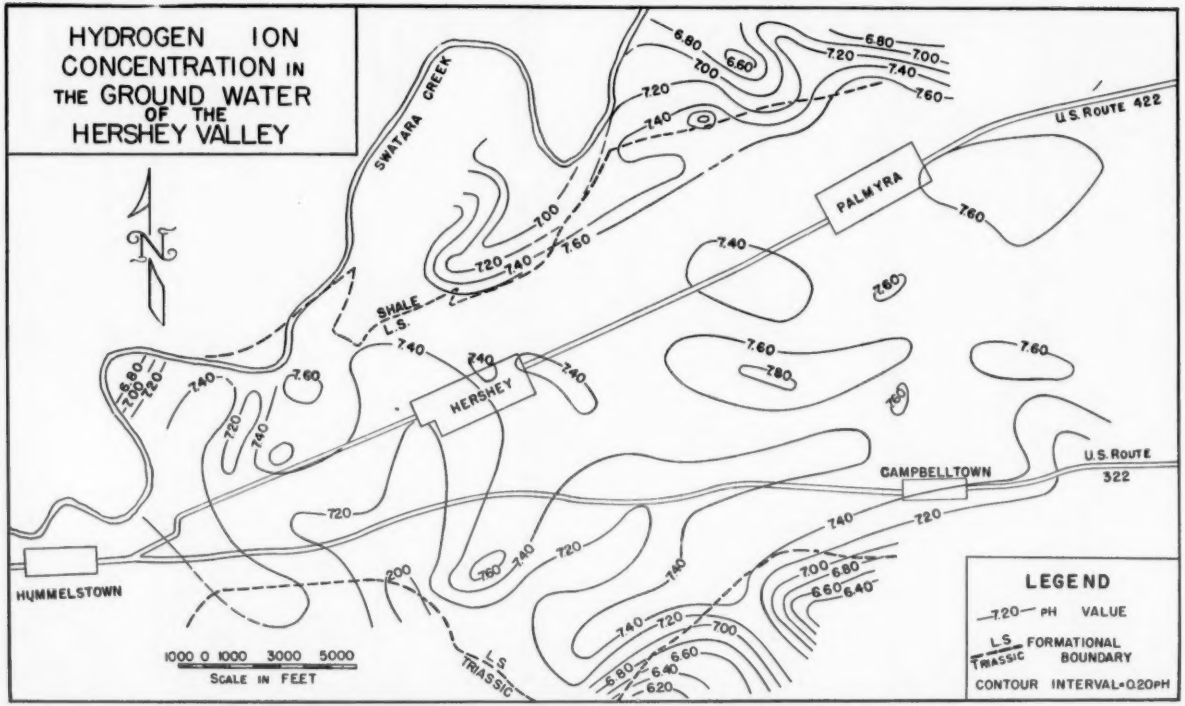


FIGURE 10. HYDROGEN ION CONCENTRATION IN GROUND WATER OF THE HERSHEY VALLEY

FIGURE 10. HYDROGEN ION CONCENTRATION IN GROUND WATER OF THE HERSHEY VALLEY

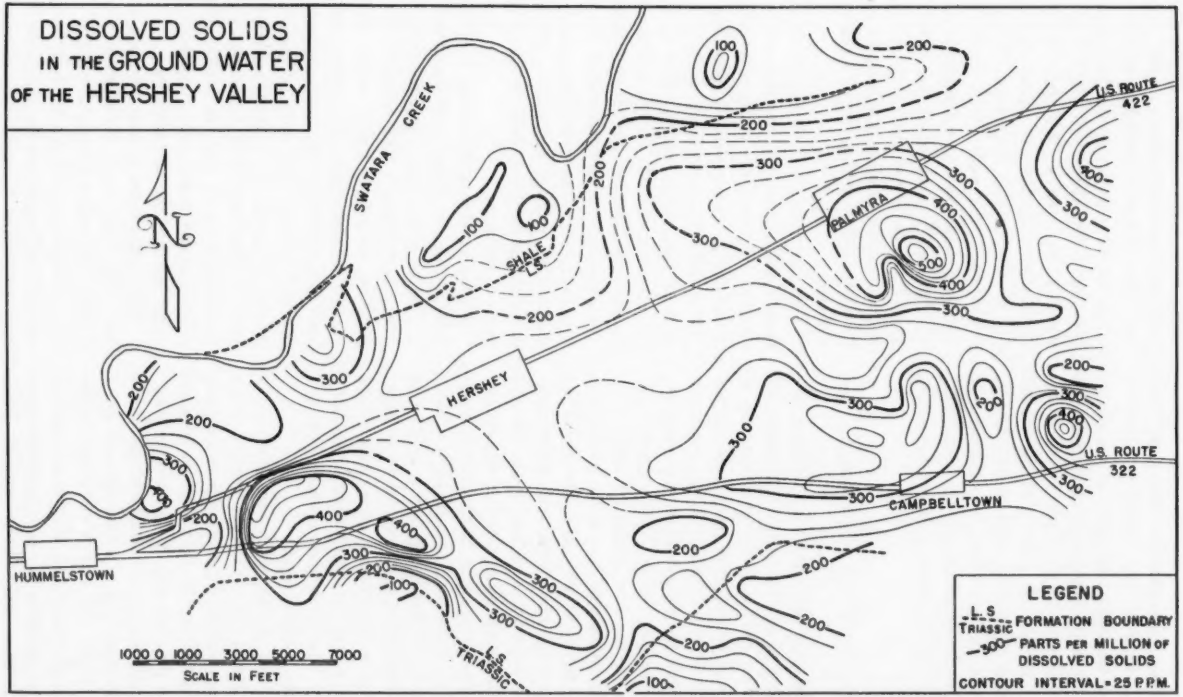


FIGURE 11. DISSOLVED SOLIDS IN GROUND-WATER OF THE HERSHEY VALLEY

studies are essential to an understanding of the changes.

Chemical-ion concentration.—Parts per million of Ca, Mg, PO_4, SiO_2, Al_2O_3, SO_4, and Fe ions were determined and carefully correlated with all other data. Many interesting relations are suggested, most of them pertaining to the ease and direction of movement of ground water in rocks of varying composition. Results are still inconclusive. These investigations in the field of hydro-geochemistry are continuing, and a report may be made on the studies within a year or two.

Ground-Water Recharge

The water table throughout the Hershey Valley was falling so rapidly during September and early October 1949 and the development of sink holes was taking place in an area of such rapidly increasing size that it seemed imperative to stabilize the ground-water surface immediately. Recharging water to the ground was thought to be the best means of doing this. Two important factors had to be considered:

Chemical purity of the water was one of the factors. Whereas sandstone and shale rocks allow water to percolate rather slowly through the intergranular openings which serve partially as a filter, limestone rocks either exclude water or allow it to move with ease through tubular solution openings, providing little or no filtering effect. It is important, therefore, never to charge impure water to the ground in an area where the ground waters are being used industrially or privately. The entire ground-water reservoir could become contaminated.

Temperature of the water was the second important factor. Much of the ground water was used industrially for cooling purposes, and the temperature of the water as initially received from the well was of prime importance. Water charged to the ground with temperatures much higher than that of normal ground water (about 52°F. in the Hershey Valley) would raise the temperature of all the ground water by mixing. If the temperature of the ground water was raised, then much more ground water would be required for cooling needs. Conversely, if water of lower temperature should be charged to the ground, it would lower the average ground-water temperature.

All water subsequently recharged was chemically suitable for any use. It was decided to recharge experimentally a small volume of water for an 8-week period to determine what effect the temperature would have on the Corporation wells. Warm (70°–76°F.) spray pond water at the rate of 1000 gpm was recharged at the Derry Spring well about 1000 feet east of the Corporation buildings. During the 8-week period, the wells closest to the point of recharge showed a water temperature increase of 1°–6° and a very small rise in water level (Fig. 12). The Corporation wells farthest west showed neither temperature increase nor water level rise, although their water levels no longer declined.

This experiment proved what had previously been demonstrated by the fluorescein tests—namely, that from the point of recharge water would move outward in all directions subject to the major factors controlling the movement, such as size, shape, and number of openings; the existence of mud "seals"; and of course the regional gradient of the ground-water table. The experiment also corroborated the expectation that ground-water temperatures would be changed by mixing, and that a gradual temperature gradient extended outward from the point of recharge.

Beginning at noon December 12, 1949, about 3500 gpm of low-temperature water was charged to the old pond area adjacent to the dried-up Derry Spring about 1000 feet east of the Corporation buildings. These waters were from various sources; a large percentage had already been used for cooling purposes and returned to the spray pond. This was the beginning of a general program for recharging low-temperature, non-contaminated waters to the ground in amounts up to 8000 to 10,000 gpm in an effort to stabilize or to raise, if possible, ground-water levels in the affected area.

The results were almost immediately beneficial. Figure 12 shows the effect on water temperature and water level in five of the Corporation wells and in Well No. 68, located 6000 feet east of the Corporation buildings and 5000 feet southeast of the Annville Stone Company. Well No. 68 had suffered a serious decline in water level during the fall of 1949. The resulting rise of the ground-water table was shown by many other wells throughout the entire valley, in-

FIGURE 12. EFFECT OF GROUND WATER RECHARGE ON HERSHEY WELLS
(Location of wells on Fig. 1)

cluding Wells Nos. 315, 415, and 501 (Fig. 4). These results demonstrated that recharged waters moved outward in many directions, even though they were under the constant influence of the steep regional gradient on the ground-water table sloping toward the Annville Stone Company. Undoubtedly there was a local "hump" on the ground-water table at the recharge point, which partly explains how the water would move opposite to the regional gradient.

LITIGATION: During January and February 1950, the Annville Stone Company and Hershey Chocolate Corporation were actively engaged in litigation growing out of the abnormal ground-water conditions in the Hershey Valley and specifically from the program of ground-water recharge practiced by the Hershey Chocolate Corporation. The Annville Stone Co. sought an injunction against the Chocolate Corporation, enjoining them to cease recharging water to the ground. The program of ground-water recharge was successfully defended, twice in the Dauphin County court and finally before the Supreme Court of Pennsylvania in Philadelphia. The suit against the Hershey Chocolate Corporation was withdrawn by the Annville Stone Company in May 1950.

RETURN TO NORMAL CONDITIONS: To reduce the burden of increasingly heavy pumping demands, which had risen from 6500 gpm to more than 8000 gpm, the Annville Stone Company engaged in a program of grouting in February and March 1950 to seal off the openings carrying water into their mine. The apparent success of this action, in addition to the recharge program, was demonstrated by a return to nearly normal conditions throughout the Hershey Valley beginning in late March 1950. Springs began to flow again. Water levels rose in wells that could once more be pumped. Spring Creek began to flow again. Sink holes no longer developed. The water-table map of May 1950 showed separate areas of ground-water depression around the Chocolate Corporation and the Stone Company mine once again. They were similar in size, shape, and location to those outlined in October 1948 (Fig. 3).

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TIME-STRATIGRAPHY AND THE MIOCENE EPOCH

BY MARTIN F. GLAESSNER

ABSTRACT

Stratigraphic geology operates concurrently with two different time scales. One is absolute and based on measurements in units of equal duration (years), the other is relative and based on arbitrarily selected and unequally spaced fixed time markers of known or unknown position in the absolute scale, established as a discontinuous sequence of unrelated events which have left traces in the geological record. These events which may be classed as lithogenetic, geotectonic, or biological, serve as fixed points for the establishment of a hierarchic system of arbitrary divisions of geological time. The first regional appearance of a new fauna, analyzed from the viewpoint of evolution as well as ecologically, is the basic time-marking event. Major subdivisions are marked by horizons at which a number of events occur more or less contemporaneously, indicating major changes. Stratigraphic nomenclature expresses the resulting classification. The principle of priority is applied to it with certain limitations.

On the basis of these principles, it is found that the existing discrepancies in the placing of the lower and upper boundaries of the Miocene can only be resolved by agreement, as various arguments for different placings are inconclusive. Ambiguity of terminology has further confused the Miocene-Pliocene boundary problem. In the Caribbean region, the terms Lower, Middle, and Upper Miocene seem to have acquired an agreed meaning. Faunal relations with the Indo-Pacific Miocene seem at first glance confusing, but a suggested line of analysis leads to the conclusion that discrepancies are small. The Indo-Pacific "letter-classification" is reviewed in some detail and rejected for regional or general use outside Indonesia as seriously defective. The introduction of Lower, Middle, and Upper Miocene as divisions of the standard time-stratigraphic scale is advocated. Most authors using at present European Stage terms for reference place the Burdigalian in the Lower, the Vindobonian in the Middle, and the Sarmatian in the Upper Miocene. There is no agreement about the desirability of including the Aquitanian in the Lower Miocene and the post-Sarmatian interval, which is currently but ambiguously termed Pontian, in the Upper Miocene.

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INTRODUCTION

Geology as a historical science requires a standard time scale to which events in the earth's history can be related. At the present time, much work is directed towards the construction of an absolute time scale, in years (or Millions of years), i.e. in units of equal, though arbitrary, length (Simpson, 1947; Knopf, 1949). This is known as absolute geochronology. While this work is in progress, geology continues to operate, as it has operated since the very beginning of geological studies, with a relative time scale, produced by an arbitrary division into unequal parts of the continuum of geological time. This is the field of relative geochronology or time-stratigraphy. The conventional relative scale is used in the construction of the absolute scale as a means of relating the necessarily scanty absolute age determinations to each other, but it seems likely to be of permanent use, depending upon clari-

fication of the nature of the arbitrary divisions and on perfecting methods of their recognition and distinction.

The only essential basis of the relative or stratigraphic scale is an established irreversible sequence of *events*. The application of this scale is comparable, in the field of human history and personal life, with the use of terms such as "during the Middle Ages", "after the French Revolution", "in the early Victorian period", "before the last war". These statements are as definite as they are intended to be and serve the purpose of relating facts of the past to each other ("Synchronology", Hedberg 1948). Similarly, statements in terms of the current stratigraphic system have a definite meaning and their inherent vagueness is not a disability which should be removed by sharper definition in terms of the absolute time scale but is to some extent intentional. What should be remedied is ambiguity of terminology.

The relative time scale in earth history differs from its counterpart in human history in its hierarchic composition. The Middle Ages are not generally divided into Periods and Epochs. A subdivision of the Eras in stratigraphy (Paleozoic, Mesozoic, Cenozoic) is well established and indispensable. This creates the problem of fitting together the subordinate divisions so that they will fill the larger units without gaps or overlaps. As time has no geographic limits, it becomes constantly more urgent to make all or some of the divisions of time world-wide so that geological events in different parts of the earth can be related to each other and their rhythms and possibly distant causes investigated. The development of stratigraphic techniques, particularly in the direction of increasing efficiency of stratigraphic correlation, requires careful attention. The much-discussed zone concept belongs to this field rather than to the formal system of time-stratigraphy.

Wider aspects of these classifications and techniques have been discussed by Hedberg (1948; 1951) who refers to system, series, stage, etc. as time-stratigraphic divisions and to period, epoch, age, etc. as time units which he does not consider as stratigraphic units because they are not material bodies. The divisions of geologic time in relative geochronology are established by stratigraphic methods, stratigraphy being more than just "description of strata", as implied by its name. This justifies the use of the convenient term "time-stratigraphic" for the time-units mentioned above, while systems, series, etc. are referred to as time-rock terms, as proposed by the American Commission of Stratigraphic Nomenclature in 1947 and in the Australian Code of Stratigraphic Nomenclature. A further discussion of Hedberg's important contributions to the concept of "stage" would be out of place in this paper which deals mainly with the Miocene as an epoch and the subdivisions of this time interval.

Since the preceding paragraph was written, the American Commission on Stratigraphic Nomenclature (H. D. Hedberg, Chairman) has published a report (1952, p. 1627–1638) in which it adopts Hedberg's terminology. To avoid confusion I have to state expressly that throughout this paper and in its title I use the term "time-stratigraphy" not in the sense now given to it by the Commission, but with reference to "geologic-time units" as defined in the new report.

Modern theories on petrogenesis, geotectonics, paleo-biogeography, and evolution use much implied stratigraphic correlation, and their authors often take age relations for granted without examining critically the basis of crucial statements. Some stratigraphers, maintain an attitude of agnosticism towards the possibility of finding a common basis of world-wide correlation, while others voice an over-critical demand for an unlimited refinement culminating in the ultimate replacement of the relative by the absolute time scale. Others again see the solution in the creation of a perfect hierarchic system of stratigraphic nomenclature in which each name is based on priority of usage and weighed down with a formal definition, irrespective of the practical requirements of the stratigrapher and the ever-increasing amount of factual observations which should be classified in terms of the continuum of time.

With Pia (1929; 1930; 1937), we see time as the primary concept in stratigraphic classification and events of various types as time markers. These events must be definable and veri-

fiable. The only verifiable events in stratigraphic geology are expressed within series of strata, either in their lithology, spatial relations, or fossil content. *Changes in lithology* were the first observed stratigraphic facts to be related to time, hence names such as "Carboniferous" or "Cretaceous". Under the impact of the developing facies concept (which in its most general form states that rocks of different composition are formed contemporaneously in co-existent different environments), the use of lithological differences in time stratigraphy lost favour. *Transgressions and regressions* of the sea have long been considered important stratigraphic time markers, but this line of argument depends far more on local evidence than that of the paleontologist. Transgressions and unconformities cannot be identified in distant areas in the manner of index fossils, and local sequences of such events are not easily combined into regional sequences, which would be a necessary step in the formation of a time scale. *Orogenic phases* have been widely used in accordance with Stille's concept of short duration and wide effects of such phases. Pia (1930, p. 74) has criticized the application of this theory, which depends for proofs very largely on conventional bio-stratigraphic correlations. A recent discussion in Geol. Rundschau (vol. 38, H. 2, 1950) has again shown the difficulty of distinguishing between stratigraphically brief orogenies and protracted epeirogenic processes. This makes it difficult to accept orogenies as time-markers, even if the underlying concepts are considered as sound.

In *biostratigraphy* several approaches to time-stratigraphic classification are open. The oldest is the gross or bulk comparison of faunal composition viewed as changing continuously or gradually towards a set of living regional standards. Lyell's percentage method, which has been much discussed lately (Finlay and Marwick, 1940; Keen, 1940; Glaessner, 1943), has given us the terms Eocene, Oligocene, Miocene, and Pliocene. Time stratigraphic conclusions can be based on evolution as shown by first appearance, change, or extinction of phyletic lines or stages in a bioseries. Morphological or statistical analysis may be applied, such as distinguishes the work on larger foraminifera by Tan Sin Hok as a major contribution to stratigraphy, or the approach may be purely empirical as in most current work in which smaller foraminifera or mollusca are used as time markers. In the stratigraphic evaluation of vertebrates, particularly Tertiary mammals, phylogenetic studies are closely integrated with the examination of climatic and other environmental influences on dispersal. Marine mammals are particularly valuable in Tertiary correlation as they make it possible to apply the refined methods of vertebrate paleontology to the age classification of marine strata. Modern workers on Tertiary marine faunas (Kautsky, Schenck, Fleming) take climatic influences on migration of faunal element into consideration. The first regional appearance of a new fauna, analyzed from the viewpoint of evolution as well as ecology, is the basic time-marking event in stratigraphy. Other events in other regions are dated as contemporaneous, earlier, or later by standard methods of correlation and may be used as reference points in subsidiary regional standards.

To stratigraphic nomenclature, as to all naming, the rule of priority applies in the sense that names, once given, are not to be changed or replaced needlessly. It is not fundamental and has to give way under certain circumstances. One is the discovery of new facts which make possible a new and clearer definition than that under which the term was first proposed. Another is the convenience of present usage, which should not be disturbed or thrown into confusion on account of priority alone.

A selected locality at which the time-marking event can be established becomes the type locality of the named time unit. Significant events may be related in different ways to the named time intervals. They may have duration, as for example a molluscan fauna of a certain composition (expressed as a percentage of living species or otherwise), which may exist in a series of strata at a certain (type) locality. More frequently, they are represented by horizons, i.e. boundary planes without thickness, such as the extinction of a species or genus or the first appearance of a species, a fauna, or a biocharacter in a bioseries. The beginning of a transgression represents a time plane wherever it is visible, but it is not likely

to be everywhere the same horizon. If the bundle of time planes represented by it is sufficiently closely spaced to be below the limits of our stratigraphic resolving power, the beginning of the transgression, although spread through a certain amount of geological time, may still be a useful time-marking event. The same reasoning applies to the geographical spreading of a distinctive fauna.

One of the main problems in time-stratigraphic classification is the selection of *significant* events as markers, that is, of horizons of *major* change where a number of important events occur contemporaneously. Again, they will be only relatively contemporaneous, within the limits of the resolving power of our stratigraphic methods. Many time-stratigraphic categories are based on long-standing usage rather than sharp definition. From a practical point of view, it is preferable to give them sharper definition by the careful study of significant events at their conventional boundaries, which are for historical reasons often primarily rock-stratigraphic rather than time-stratigraphic horizons. It will be of little practical value to select and propose an entirely new set of time markers, however well defined, as few geologists will be willing to replace or alter fundamentally the traditional time scale.

The following discussion concerns the present status of the Miocene and its subdivisions as units in this time scale. It is not possible, within the limits of available space, to review the nature of accepted time-marking events and the observations which have led to their recognition. The application of the principles discussed in the preceding paragraphs to the currently accepted time scale as reviewed here is expected to assist materially in solving the remaining problems.

What Is "Miocene"?

The term Miocene was introduced by Lyell (1833, see Schenck 1935, p. 532). He referred at first to the beds of the Superga Hills near Turin in Northern Italy as "the type", and named as examples Miocene formations in the Touraine (which he later just as definitely considered as "the type", e.g. in "The Student's Elements of Geology", 2nd ed., 1874, p. 192), in the vicinity of Bordeaux, Piedmont, and the basin of Vienna. The following epoch, the Pliocene of Lyell, has as its type "the subapennine strata of Northern Italy" which has come to be interpreted as including the Stages Plaisancian and Astian. Thus at least the European Stages Burdigalian and Vindobonian (which comprises the Helvetian and Tortonian) and their time equivalents are definitely included in the Miocene. To this extent, the concept of the Miocene Epoch has an agreed meaning which is not affected by any ambiguity concerning its upper and lower limits.

There is hardly any area containing a major development of Late Tertiary strata where the term Miocene is not in current use. In the historical development of the study of faunas from remote areas such as Indonesia, Australia, New Zealand, or Patagonia, European affinities are stressed in early stages. Later, attention is given to the provincial characters and, as a result, the applicability of the standard European time scale is often denied. As investigations progress, links with areas which are more firmly placed in the standard scale are re-established on a solid basis, either by the discovery of immigrant forms of wide geographic range or step by step by a process of linking through adjacent regions. The margin of error involved in the application of the well-known techniques of bio-stratigraphic correlation becomes increasingly serious as they are applied to minor time divisons.

Oligocene-Miocene Boundary

There are no objective criteria by which the lower and upper limits of the Miocene are defined. The establishment of limiting events for the Aquitanian Stage does not prove whether it is Miocene or not. This difficulty is caused by the hierarchic nature of the relative stratigraphic scale. The historical approach fails to provide the answer because the standard stratigraphic scale grew by intercalation of new units to fit newly discovered facts. It does not help to refer to the definitions of Lyell, who recognised only Eocene and Miocene for a distinction between Miocene and Oligocene. Naturally, events in the "type area" are particularly important, but in the case of

the boundaries of the Miocene, we are dealing with three different type areas. The four-fold division of the Tertiary suggested by new discoveries in the middle of the last century proved more convenient than the three-fold division. This makes it necessary to reach agreement about the redistribution of strata and faunas among the major units (series). It is interesting though not decisive that Beyrich placed at the top of his Oligocene the Sternberg sands of northern Germany (Dehm, 1949), as it would be interesting to know definitely whether the base of the Superga sequence includes Aquitanian equivalents (Schenck, 1935). We may suppose, for the sake of argument, that the Sternberg sands are Chattian and that the type Aquitanian (Rutsch, 1952) is entirely younger, as shown by Brinkmann (1948, table facing p. 256) but denied earlier by Oppenheim (1913) and Kautsky (1925, p. 25). Supposing further that the Superga sequence has no Aquitanian but starts with a Burdigalian conglomerate, we would have to exclude the Aquitanian from both the Oligocene and Miocene as originally defined (this was done by Stirton, 1951). The possibility of such a conclusion indicates clearly that stratigraphic classification cannot be based on priorities, original definitions, and other purely historical considerations. The failure of this approach leaves faunal and stratigraphic arguments to be considered. The arguments have been ably and dispassionately reviewed by Gignoux (1950, p. 508–10, 598; see also Schenck, 1935). A number of prominent paleontologists have argued in favor of including the Aquitanian with the Oligocene (dal Piaz, 1929, Denizot, 1940; Hürzeler, 1946) while others have placed it in the Miocene (Dehm, Finlay, Simpson, Tan). Taken separately, these arguments cannot produce any more decisive result than those based on stratigraphic relations in type areas or on historical considerations. The necessity of a distinction between Oligocene and Miocene is unquestioned and these Epochs (Series) are "natural" divisions as long as we are some distance (in time or space) from their boundary. It seems that it will have to be defined by agreement, by a representative body or commission which will consider all arguments which have been put forward.

Miocene-Pliocene Boundary

The upper limit of the Miocene presents similar problems which will have to be solved in a similar manner, but they are further confused by ambiguous terminology. The time-stratigraphic terms most frequently used for Late Miocene or Early Pliocene are Sarmatian and Pontian. The name Sarmatian was introduced in the following terms by E. Suess in 1866 (translation of a quotation from Papp and Thenius, 1949, p. 770):

> "To give them a comprehensive name I shall designate in future (with the concurrence of M. Barbot de Marny who has contributed so much to our knowledge of their eastward extension) all these deposits, i.e. our *Cerithium* beds and the Hernals clay, as the Sarmatian Stage, and that eastern fauna which includes *Mactra podolica*, *Donax lucida*, etc., as the Sarmatian fauna".

This shows clearly that E. Suess is the author and the Vienna Basin is the type area of this Stage. Two important conclusions follow. Firstly, the Sarmatian is younger than the Vindobonian, a term designating Suess' "Second Mediterranean Stage" which underlies the Sarmatian everywhere and contains a fauna differing fundamentally from the characteristic "eastern" fauna discussed by Suess. As the Tortonian age of the younger and the Helvetian age of the older of the faunal assemblages included in Suess' Second Mediterranean Stage has been corroborated by recent detailed faunal comparison (Janoschek, 1951); we cannot follow Gignoux (1950, p. 600–602) who considers the Sarmatian as a brackish facies of the Tortonian. Secondly, it is wrong to consider Barbot de Marny as the author of the term Sarmatian and Southern Russia as the type area. This claim is based mainly on the derivation of the name, which was selected apparently to draw attention to the supposed eastern origin of the fauna while the typical strata were clearly indicated by their local names.

Laskarev, N. Andrusov, and D. Andrusov have shown (D. Andrusov, 1932; 1938) that the application of the term Sarmatian east of the Carpathians was extended upward beyond the limits of the stratigraphic unit for which it was introduced. In the Vienna Basin, the beds containing the Sarmatian fauna are followed by the "*Congeria*-Stage" which soon came to

TABLE 1. NOMENCLATURE OF EUROPEAN LATE MIOCENE AND EARLY PLIOCENE STAGES

Ponto-Caspian Basin	Pannonian Basin		Vienna Basin	Italy	Spain	
Dacian* = Kimmerian*				?	Daciense	Pontiense
Pontian*	Upper	Pannonian*	Paludina Sands	Pontian	Pontiense Superior (P. sensu stricto)	
Meotian*	Middle		Congeria Beds		Meotico	
?	Lower					
Upper Sarmatian = Chersonian*					Vindobo-niense Superior	Vindobo-niense
Middle Sarmatian = Bessarabian*						
Lower Sarmatian = Volhynian*			Sarmatian*	Sarmatian		
			Vindobonian* = "Second Mediterranean"	Tortonian*	Vindobo-niense Inferior	
				Helvetian		

* Names in type areas.

be considered an equivalent of the Pontian. This term goes back to the introduction by Le Play in 1842 of the "Formation tertiaire pontique" for the limestones of Odessa and Novocherkassk (now included in the Pontian s.str.) and of Taganrog (now known to be Sarmatian) in Southern Russia. Barbot de Marny restricted the term Pontian in 1869 to exclude the Sarmatian, which was divided in 1899 by N. Andrussov into Lower, Middle, and Upper Sarmatian. These parts were named Volhynian, Bessarabian, and Chersonian, respectively, by Simionescu in 1903. It has been shown that only the Volhynian and part of the Bessarabian correspond to the Sarmatian in the Vienna Basin. The succeeding strata here and in the adjoining Pannonian Basin (Hungary) were named Pannonian by Roth von Telegd in 1879, while N. Andrussov in 1890 intercalated the Meotian between his "Upper Sarmatian" and the restricted Pontian in Southern Russia. The *Congeria*-beds of the Vienna Basin are therefore not Pontian s.str. but correspond to the lower and middle Pannonian of Hungary and the Chersonian and Meotian of the Ponto-Caspian Basin; the Pontian s.str. is represented in the Vienna Basin by the "*Paludina*-sands" (Upper Pannonian).

Thus we find the Pontian being raised and restricted in the east while it became expanded and its base lowered in western Europe. This western European usage, which conflicts with observed facts in the type area, has been adopted in many inter-regional correlations as part of the standard time scale. It confuses comparisons between "pre-Pontian" faunas in different areas which are not contemporaneous (Sarmatian *sensu stricto* in the west, Chersonian or Meotian in the east). The term Sahelian, established by Pomel in 1858, has been used for the Late Miocene or Mio-Pliocene in a more purely marine facies. Recent work by Laffitte (see Gignoux, 1950, p. 602) has shown, that the Sahelian of Algeria is Tortonian

separated by a disconformity from the overlying Pliocene. The Sahelian of Tunisia is Pliocene. Gignoux concludes that the Sahelian has no paleontological or stratigraphic "individuality" and that the term should be discarded.

A recent paper by Crusafont Pairó (1951) gives a clearer account of stratigraphic and faunal relations in this part of the Tertiary sequence of Western Europe (Spain), but the author has not yet introduced an acceptable terminology. He found marine, brackish, and continental faunas closely associated throughout the interval from Burdigalian to Pliocene. A brackish fauna of Sarmatian type is intercalated in a sequence containing a marine fauna of Vindobonian character. This appears to be the previously unknown marine equivalent of the Eastern European Sarmatian (*sensu lato*). The suggested use of "Upper Vindobonian" for this marine development extends this term to include the Sarmatian and part of the Pannonian which are certainly not the upper Vindobonian of the Vienna Basin (Table 1). The succeeding strata contain a vertebrate fauna transitional to that of the next higher "true" Pontian and are correlated with the Meotian. At the same time, these two upper Stages are combined as Pontian *sensu lato*, to follow upon the Vindobonian *sensu lato*, in accordance with the French usage. The Spanish Meotian seems to agree in fauna and stratigraphic position with part of the Lower Pannonian of the Vienna and Hungarian Basins. Crusafont Pairó expresses himself cautiously on the position of the Miocene-Pliocene boundary.

This boundary is drawn between Sarmatian and Pannonian in the Vienna Basin, between Meotian and Pontian *sensu stricto* by most Russian authors, and between Pontian *sensu lato* and Plaisancian by French stratigraphers. The currently favored definition of this boundary seems to rest on the appearance of the *Hipparion*-fauna (Koenigswald, 1939; Stirton, 1939). Its earliest occurrence is in the Chersonian, and it has been calculated that its spread to Western Europe could not have caused a stratigraphically noticeable time lag (Papp and Thenius, 1949). As the Chersonian is younger than the typical Sarmatian, the occurrences reported as "Upper Sarmatian" do not affect the position between the typical Sarmatian and the Pannonian of the Miocene-Pliocene boundary *as based on vertebrate faunas*. (*See also* Colbert, 1942, p. 1439 ff.). Here again, arguments based on vertebrate faunas only have no more absolute validity than the vertebrate-paleontological arguments for placing the Aquitanian in the Oligocene. Reviewing current opinions from a regional point of view, we find in many correlations of American Tertiary strata (Cooke, Gardner, and Woodring 1943; Weaver *et al.*, 1944; Senn, 1940) the Pontian placed in the Upper Miocene, as it is by Glaessner (1943), Oostingh (in van Bemmelen, 1949, p. 94), and Finlay (1947) for the Indo-Pacific Region, while it is retained in the Pliocene by van Bemmelen. Most Central and Eastern European authors consider it as Pliocene. Gignoux (1950, p. 602) says that it straddles the Miocene-Pliocene boundary. This is often expressed by the ambiguous term Mio-Pliocene, corresponding to the equally objectionable term Oligo-Miocene. The use of such terms proves that the boundaries are not satisfactorily placed in relation to recognizable age divisions and that a decision by agreement is required.

Caribbean and Tropical Pacific Miocene Correlation

A province where intensive regional correlation of Tertiary deposits has produced important results in time-stratigraphic classification is the Caribbean region. Local problems still exist, but a number of tabulations and discussion based on detailed local zoning show that a very large measure of agreement has been reached (H. H. Renz 1942; 1948; Senn, 1940; Stainforth, 1948a; 1948b; Cooke, Gardner, and Woodring, 1943). Zoning by larger and smaller foraminifera plays the most important part in these correlations. Age classifications are given in terms of a division of the Tertiary into Epochs and of the Miocene (which alone is here discussed) into Lower, Middle, and Upper Miocene. Comparing these divisions with the European Stages, all authors include the Burdigalian in the Lower, the Helvetian and Tortonian in the Middle, and the Sarmatian and Pontian in the Upper Miocene, but Senn

uses the term Pontian *sensu lato* including the Sarmatian, and Cooke *et al.* still use Sahelian for both. The only disagreement in general time-classification concerns the Aquitanian, which Cooke *et al.* place in the Miocene while all other authors mentioned consider it as Upper Oligocene. The difficulties and uncertainties of trans-Atlantic correlation are stressed wherever European Stage terms are mentioned, and there is some conflicting evidence, particularly from Pteropods which have Burdigalian affinities in beds placed on other evidence in the Upper Oligocene. Paleogeographic changes in Late Middle and Upper Miocene make foraminiferal correlations difficult. Owing to widespread regression, many faunas have a brackish character and become poor in species while others, formed in separated basins, are too provincial in composition to be stratigraphically useful (H. H. Renz, 1948; Stainforth, 1948a).

Faunal relations between the Indo-Pacific Miocene and tropical America are peculiar and somewhat misleading to casual observers, who have pointed out similarities between the Caribbean Oligocene faunas and the Miocene of the Western Pacific. In the Indo-Pacific Region, we know at present the distribution of smaller foraminifera from about the base of the Miocene upward. In tropical America, we know it well from Eocene to Middle Miocene. Under these circumstances, similarities of heterochronous faunas can be expected and are to be interpreted as evidence of an endemically evolving tropical Pacific fauna which cannot be traced back to its origins in the west or to its extinction or its Late Tertiary descendants in America. The endemic forms are associated in America with elements which never reached the Western Pacific (many *Siphogenerina*, *Valvulineria*, the *Bulimina sculptilis* line) while some common Indo-Pacific foraminifera are probably of European origin (*Eponides praecinctus*, *Nummoloculina*). The local origin of others can be traced within the Miocene by transitions (*Pulleniatina* from a *Globigerina* in Middle Miocene, *Bolivinita quadrilatera* from *Bolivina compressa* in Lower Miocene). Some genera such as *Quinqueloculina* and *Robulus* seem to contain groups of different origin. *Orbulina* occurs rarely in the Middle Oligocene of Venezuela and Porto Rico (H. H. Renz, 1948) and then commonly in the Late Oligocene (probably Aquitanian) of the Caribbean Region (Renz, Stainforth) and in the Lower Miocene (Burdigalian) of Australia and New Zealand. Le Roy's "*Orbulina*-surface" in Sumatra (1948; 1952, criticized by Brönnimann 1951, p. 134) occurs in beds generally placed in the Aquitanian to Burdigalian. It seems that the remaining discrepancies in trans-Pacific correlation are small, probably of less than Stage (Age) magnitude.

Letter Classification in Indonesia

The regional "letter classification" of the Indonesian Miocene requires more detailed consideration, particularly in view of its attempted application in the Australian region. I have previously reviewed its development up to about 1939 as far as it affects correlation (Glaessner, 1943).

Stage e.—There is evidence of a paleontological correlation of the top of the e-stage with the top of the Aquitanian (Tan, 1906; 1939). A slight divergence from this view is found in Mohler's correlations (1949) which were accepted by van Bemmelen (1949). Mohler considers the uppermost portion of the e-stage (Substage e_3) as the lower part of the Burdigalian. The only reason given for this correlation is the occurrence of *Flosculinella reicheli* Mohler. The supporting reference to Reichel's monograph of the Alveolinidae proves nothing, as *Flosculinella* was known to this author only as a genus represented by two species from beds in Indonesia which were placed in the Burdigalian on the evidence of the accompanying fauna. Its discovery in earlier beds containing the *Eulepidina-Spiroclypeus* fauna means only that *Flosculinella* existed somewhat earlier than previously known, but not that the Burdigalian must now be extended downward. As M. Rutten has pointed out in his synopsis of the letter classification (in van Bemmelen, 1949, p. 83–88), there cannot be an Alveolinellidae stratigraphy taking precedence over other stratigraphic systems, but only a faunal stratigraphic classification and correlation.

A subdivision of the e-Stage was attempted, on faunal grounds, by Leupold and van der Vlerk (1931) who introduced "zones" e_1 to e_3

without presenting detailed stratigraphic evidence of the vertical distribution of critical forms in measured sections. It was rejected later by Tan and Mohler. Mohler (1949) accepts only a division between e_{1-4} and e_5, indicating in a chart the last appearance of *Neoalveolina (recte Borelis)* and *Eulepidina formosa* and the first appearance of *Flosculinella*, *Miogypsina (s.str.)*, *Miogypsinoides dehaarti*, and two species of *Lepidocyclina* as distinguishing features. There is no evidence at present to show which part of the e-Stage is Chattian and which is Aquitanian, but Schenck and Childs' work (1942) indicates that this distinction can be based on the top range of *Lepidocyclina s.str.* when more representatives of this subgenus are found in definable stratigraphic positions. The difficulty of defining the base of the Aquitanian in Indonesia is an argument in favor of its inclusion in the Oligocene.

STAGE f.—The f-Stage was divided by Leupold and van der Vlerk into three "zones". Their criteria can no longer be used, as Tan Sin Hok's work has extended the ranges of many of the larger foraminifera considered diagnostic. Tan accepted two divisions of the "Middle Neogene" corresponding to the Stages n_2 and n_3 established by de Jongh in 1933 as a local standard for the work of the Netherlands Indies Geological Survey. In 1943 I correlated these divisions with the European Stages Burdigalian and Vindobonian, Oostingh's Javanese molluscan stages Rembangian and Preangerian, and Leupold and van der Vlerk's "zones" f_{1-2} and f_3. Van Bemmelen (1949) places some (p. 108) or all (p. 103) of "zone" f_3 in n_2. The criteria given by Tan are not fully reliable, and they do not coincide with the distinguishing characters of f_{1-2} and f_3. This becomes clear when a list of all known species of larger foraminifera from strata originally included by Leupold and van der Vlerk in f_3 is compiled (Tji Lanang, Njalindoeng, Gelingseh, Menkrawit, and Lower Palembang beds). This list includes twenty species, among them *Lepidocyclina sumatrensis*, *L. borneensis*, *L. angulosa*, *L. inflata*, *Miogypsina bifida*, *Katacycloclypeus annulatus*, and *Flosculinella bontangensis*. Most of these should be restricted to the lower division of the f-Stage. On the other hand, *L. rutteni* occurs in the Telisa beds which are definitely older than f_3. Only *L. orientalis*, *L. eulepidinacea* (*syn.: L. talahabensis*), and *L. gigantea* have not been recorded from older strata. Van Bemmelen's compilation of stratigraphic data shows that, at a number of localities, species considered by Tan as diagnostic for n_2 or n_3 occur together.

It is however possible, as recognised by Mohler (1950), to distinguish a lower division with *Flosculinella bontangensis* (restricted), *F. globulosa*, *Austrotrillina howchini* (last occurrence), *Miogypsinoides dehaarti* (last occurrence), and other species. Mohler correlates its top with the top of the Burdigalian. The extinction of *Austrotrillina* supports this correlation (Glaessner, 1951). Whether the remaining upper part of the f-Stage can be further subdivided is still uncertain. Mohler identified his lower division with Leupold and van der Vlerk's f_1, but he and M. Rutten categorically denied the possibility of a distinction between f_2 and f_3 (in van Bemmelen, 1949, p. 86, 136–7). It may yet be possible to recognise generally a zone with the last *Katacycloclypeus*, the first *Tryblio-lepidina rutteni*, common *L. ferreroi*, *L. angulosa*, etc., as in the Ojo beds of Java and equivalents of the Njalindoeng and Tji Lanang beds and possibly in New Guinea and Fiji, and make it a time-stratigraphic division. It can be distinguished from the youngest *Lepidocyclina*-bearing zone, which contains mainly *L. orientalis* and *L. gigantea* but no *L. rutteni* or *Katacycloclypeus* and is not the f_3 Stage as originally defined. Leupold and van der Vlerk's f_2–f_3 boundary, which cannot be recognized was eliminated from van der Vlerk's latest table (1950). The description of a sequence of two faunal zones as, respectively, f_{1-2} and f_{2-3} or as f_{1-2} and f_3 is meaningless at the present state of our knowledge of biostratigraphy in Indonesia, where the letter classification originated, unless it refers, with a confusing shift of terms, to the distinction between f_1 and f_{2-3} according to Mohler.

While the lowest recognizable division of the f-Stage is probably Burdigalian, there is as yet no evidence of a correspondence of the two higher divisions with the Helvetian and Tortonian of Europe. There are also good reasons against a generalization of Oostingh's Javanese mollusca Stages Rembangian and

Preangerian, which I had thought practicable in 1943. Van Bemmelen's comprehensive review shows that the faunas on which these Stages were based have not been found in clear superposition and that their mutual stratigraphic relations are uncertain. The Rembangian is placed in e_5, f_1, and f_2, which means that its top is not otherwise faunally identifiable. The molluscan stages cannot be clearly related to the regional succession of foraminiferal faunas and cannot be recognised outside their type areas. This is shown by a large fauna described by Beets (1941) from Borneo, which on molluscan evidence belongs between the Njalindoeng and Tji Lanang faunas, i.e. in the Preangerian, but which occurs on the boundary between the Taballar beds, placed on coral evidence on the Stage f_1 (Burdigalian), and the Menkrawit beds, originally considered as f_3 but now also placed in f_1 (van Bemmelen, 1949, p. 141).

Stage g.—The top of the f-Stage was originally defined as the horizon of extinction of *Lepidocyclina* and *Miogypsina*. It is followed by Stages g and h. Stage h definitely includes equivalents of the Plaisancian and Astian (van Bemmelen, 1949, p. 103, 108) while Stage g is considered to represent the Sarmatian and probably at least part of the Pontian. Distinctive faunas of smaller foraminifera charazterize these Stages in Indonesia and in New Guinea where they follow above the beds with the youngest *Lepidocyclina* (*L. gigantea*, *L. orientalis*). *L. gigantea* occurs in Java in the Bentang beds (with *L. radiata* and *Multilepidina*). These beds are now placed in the Upper Miocene g-Stage (van Bemmelen, 1949, p. 108). The discrepancy between this classification and the definition of Stage f is unexplained. Another anomalous occurrence of *Lepidocyclina rutteni f. stellata* is recorded (van Bemmelen, 1949, p. 606) from the Tapak limestone of Central Java. This contains 53 per cent recent corals and 55–59 per cent recent mollusca and is considered by Tan, Umbgrove, and van Bemmelen as Lower Pliocene. In the same general area, the correlated Kapung limestones, presumably of the same age, contain "rare *Trybliolepidina*". They are conformably overlain by the Kalibiuk beds with a Cheribonian (Plaisancian) molluscan fauna. There is even a reference to still younger *Lepidocyclina* occurrences (van Bemmelen, 1949, p. 586) in the young Pliocene (Sondean) Klitik limestones of East Java and "it seems that even in the Putjangan layers small *Lepidocyclina* do occur". These beds are Lower Pleistocene. The possibility of these fossils being derived from older strata is not discussed.

These unexplained discrepancies add to the difficulty of accepting the conventional definition of the boundary between the Stages f and g. Experience in New Guinea and other areas indicates that, at least in most sections, an interval without *Lepidocyclina* and *Miogypsina* in situ exists between the distinctive youngest *Lepidocyclina* zone and the Cheribonian faunas of the Stage h. As these are Plaisancian, the interval which corresponds to the provisional Stage g is likely to represent the Sarmatian and "Pontian sensu lato". A distinctive fauna of smaller foraminifera representing this "Upper Miocene" was described from the Kalea formation of Siberoet Island (Le Roy, 1941; van Bemmelen, 1949, p. 163–4).

It is clear that the letter classification has become defective and that no improvement can be expected in the near future, retabulations notwithstanding, without extensive detailed stratigraphic field work and paleontological investigations or publication of data which are being kept in private files. From a practical point of view, a scale of time stratigraphic terms consisting of the symbols e_{1-4}, e_5, f_1, f_{2-3}, g, h is undesirable. The alternative terms Lower, Middle, and Upper Neogene, which have been proposed for the Stages e, f, and g + h, are incorrect as at least part of the e Stage is Upper Paleogene (Oligocene), and have not found much favor. A more generalized terminology is preferable to the direct application of European Stage terms, and the use of the terms Lower, Middle, and Upper Miocene as their approximate equivalents is proposed to replaced the terms f_1, f_{2-3}, and g of the letter classification.

Generalized Scale of Miocene Stages (Ages)

The generalized time-stratigraphic divisions Lower, Middle, and Upper Miocene should become the basic units of the standard time

scale for the following reasons. The European Ages and Stages which are traditionally the basic units in the relative time and time-rock classification also serve as regional standards in Europe, and much work remains to be done there in order to give them and their constituting events clearer definition. In some instances it is not yet clear which local Stage will emerge as the unit in the regional scale. The suggested terminology cannot be conveniently developed into finer subdivisions (which would require awkward terms like Lower Upper or Upper Lower Miocene), but this is hardly necessary as regional standard units with local names will suffice for them. Present methods of correlation hardly justify world wide use of units below the rank of a Stage. The three-fold division of the Miocene is being used by many authors in different regions as a time scale for inter-regional correlation which is free from the suggestion of precise contemporaneity with deposits or faunas at a distant type locality inherent in the use of primarily local Stage names. It also obviates the arbitrary selection of one local set of names in preference to another for world-wide correlation. Yet it is free from ambiguity, as in the last 10 years or so a considerable measure of agreement has been reached about the manner in which the European Stages are to be distributed among the three divisions. The terms Lower Miocene = Burdigalian, Middle Miocene = Vindobonian (Helvetian + Tortonian), Upper Miocene = Sarmatian are being used in the Vienna Basin by Papp and Thenius (1949) and Janoschek (1951) and the following authors agree in placing the Burdigalian in the Lower Miocene: Brönnimann (1940) in Morocco, Henson (1950) in the Middle East, Colbert (1942) in India, Renz and Küpper (1947), Mohler (1949), van Bemmelen (1949, p. 108) in Indonesia, Finlay (1947) in New Zealand, Senn (1940), Cooke, Gardner, and Woodring (1943), H. H. Renz (1948) in America. The well known European textbooks by Fourmarier (1950) and Brinkmann (1948) follow the classification outlined above, with minor modifications.

Where similar agreement exists, Upper, Middle, and Lower (or Late, Middle, and Early) divisions of series and epochs should be recognized as parts of the standard stratigraphic scale, in preference to extended regional stages and local terms which now play a dual part in stratigraphic classification, bringing with them arguments and ambiguity from their type areas and a false suggestion of precise correlation.

The upper and lower limits of the Miocene, as indeed those of other Tertiary Epochs and Series, and particularly the Cretaceous-Tertiary boundary, will have to be fixed by agreement. It is suggested that a commission appointed by an International Geological Congress should receive and review all factual arguments and on that basis work out arbitrary definitions of these units. Pending a decision on this matter, the authors' views on the disputed boundaries should be made clear in publications dealing with them. For the rest, stratigraphers will do well to follow the sound advice given by Gignoux (1950), p. 599, footnote 2): "L'essentiel est d'établir d'abord des stratigraphies locales, qu'on ne généralisera pas qu'avec prudence".

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RADIATION DAMAGE IN ZIRCON: A POSSIBLE AGE METHOD

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ABSTRACT

The relationship has been studied between crystal structure damage and total alpha particle irradiation in zircon. The total irradiation was based on estimated or measured age times alpha activity, and the damage was determined by the change of the x-ray diffraction angle from the (112) plane as measured to .01° by x-ray spectrometer. The angle was found to be 35.635° (2θ) for completely undamaged material and it approached 35.1° asymptotically as the zircon had suffered increasing dosages of alpha irradiation and become metamict. The difference between the measured angle and 35.635° was found to equal $0.535 \exp [-2.31 \times 10^{-16}_{\alpha}]$ degrees, where α = total irradiation in alphas/mg. The results have been interpreted to indicate that 4.5×10^{8} atoms are displaced per alpha disintegration and that the annealing rate is essentially zero at earth's surface temperature. The process appears to be usable for age measurement but generally limited to zircons of less than 500 alphas/mg./hr. activity.

It was further found that the zircon became "half-metamict" after a radioactive energy release of 685,000 cal./gm., indicating that only a fraction of a per cent of the energy can be stored in the structure as disordered atoms, the remainder being lost as heat. Observations on the high annealing rate of common rock minerals shows that the amount of energy capable of being stored in a rock by alpha damage is negligible in questions relating to petrogenesis and volcanism.

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Introduction

Damage to crystal structures by alpha radiation, producing a metamict state, is well known in a number of radioactive minerals.[1] In this paper, the process of zircon breakdown is investigated, and also the possibility that a method of age determination can be based on this behavior.

Pabst (1952) in a presidential address to the Mineralogical Society of America reviewed and discussed the literature on the subject of metamict minerals, and the nature of the processes that cause this state, and contributed valuable new information on the crystallographic changes that take place in thorite as a result of heating. He found that thorite, which normally shows diffuse lines of tetragonal $ThSiO_4$, and which is always metamict to some degree, changed to a monoclinic form of $ThSiO_4$ after heating at 1200–1400°C. He remarked also on the fact that a monoclinic form of natural thorium silicate had been found, and that it was not in the least metamict. However, he found that isometric ThO_2 always appeared at intermediate stages in the heating, and that a mixture of ThO_2 and silica would yield monoclinic $ThSiO_4$ from a synthetic mixture. From this report it is gathered that normal metamict thorite yields isometric ThO_2 on heating to 700° and the mineral does not return to its tetragonal state as in the case of zircon; and that the formation of a monoclinic form of $ThSiO_4$ is an unrelated phenomenon which occurs when ThO_2 and silica are heated together, and that in this form the material does not become metamict. Pabst also observed rotation patterns on single crystals of metamict zircon during heat treatment. He found that the single crystal zircon pattern before heating gave way to a mixture of diffraction points from a slightly disoriented zircon structure and from cubic ZrO_2 after heating to 950°C. Further heating to about 1400°C. removed the ZrO_2 but left the zircon in the form of a complex of disoriented crystalline domains with the zircon structure. This work agrees with the careful study on the heat treatment of metamict zircon by von Stackelberg and Chudoba (1937). Mügge (1922) also observed the disoriented crystalline phases formed from single crystals of metamict gadolinite by optical methods.

Hutton (1950) discusses the nature and cause of the metamict state, and further examines the idea of "auto-oxidation" put forth by Ellsworth (1925; 1932) which was supported by Tomkeieff (1946), who plotted PbO against the relative proportions of UO_2 and UO_3 in uraninite. However, the breakdown of crystal structures by processes other than, or secondary to, alpha particle bombardment, should be considered independently in order not to introduce confusion. The mechanism of metamictization of such minerals as uraninite is therefore discussed separately at the end.

Hutton believes that oxidation does not explain such cases as monazite, titanite, and thorianite which remain unmetamict despite considerable content of radioactive elements, and he examines the possibility that the cause of metamictization lies in the stability of the crystal structure. He concludes, on the basis of radius ratio and anion-cation distances alone, that the structure of xenotime would appear to be more stable than that of zircon, perhaps explaining why xenotime does not become metamict. However, in further examining the structure of thorite, he feels that there appears to be little reason why thorite should be so readily altered and a thorium-xenotime apparently stable. Of particular interest to the discussion in this paper, is the excellent plate in Hutton's paper showing microscopic inclusions of other minerals in grains of detrital zircon.

In general, it may therefore be said that there is no unanimity of opinion regarding the process of metamictization other than that it is caused initially or promoted by alpha bombard-

[1] Among the radioactive minerals that are usually metamict are brannerite, euxenite, fergusonite, polycrase, samarskite, thorianite, thorite, pyrochlore, betafite, eschynite, delorenzite. Radioactive minerals that are not metamict are autunite, carnotite, gummite, metatorbernite, monazite, tyuyamunite, uvanite, sphene.

ment of the crystal structure. Several workers have contributed measurements on the parameters of the process that are of interest in this discussion. Bauer (1939) correlated density, hardness, optical properties, solubility, and effects of heat treatment on a number of zircons. He concluded that solubility in certain acids was one of the best ways of measuring the degree of metamictization in the mineral. Faessler (1942) measured the heat given off by gadolinite as it reverted to its original form at elevated temperature, and found that the heat of the transformation was of the same order of magnitude as the heat of crystallization of comparable material. He measured also the specific heat of reconstituted and of metamict gadolinite and found that the relations of the specific heat are similar to those of crystalline silicates and their glassy equivalent. Kerr and Holland (1951) used differential thermal analysis as a means of comparing the heat given off by metamict minerals, but used the method as a means of identification, rather than attempting a quantitative measure of the amount of heat evolved in the process of reconstitution of any structure. Previous to this, Holland and Kulp (1950) suggested the use of similar thermal curves coupled with the measurement of the radioactivity of the mineral as a method of age determination. Morgan and Auer (1941) worked with zircons from granites of different ages and showed a correlation of optical properties with radioactivity, but no simple relationship with the age of the granite. Hurley (1952) measured the helium content of a number of zircons of different age, and radioelement content, and concluded that helium leakage increases progressively with the disordering of the crystal structure. The total amount of helium increases within the mineral at the start, reaches a maximum and then decreases again. in a function involving time and alpha activity. Henderson and Sparks (1939) made careful measurements on the damage in biotite surrounding zircon grains by the density of the pleochroic halos, and correlated these measurements with the age of the rock.

In the present paper a study is reported of the damage to the crystal structure of zircon as a function of both time and alpha irradiation. The term "alpha irradiation" here includes time also, being a total dosage of alpha particle bombardment, and is given in terms of alphas per milligram of mineral. Thus time comes into consideration twice, being a parameter both in the irradiation dosage and in the annealing or restoration of the structure at temperatures near the earth's surface.

Zircon Measurements

Samples

This investigation began as a continuation of the work reported by Hurley (1952), in which alpha radiation damage was studied as part of an investigation of the helium method of age determination. Samples and alpha activity measurements were therefore inherited in large part from this previous work. After a number of x-ray measurements had been made, it became apparent that there was enough consistency in the relationship between cell dimensions and total alpha irradiation to call for a more complete study of the subject. In the course of the work a few more samples were added to the previous list, and a number of additional alpha measurements were made to improve the previous values.

Most of the Ontario zircons were kindly donated by D. F. Hewitt, of the Ontario Bureau of Mines, who has been studying the pegmatite localities of this province. All of them are from pegmatites. Many of the remaining samples were generously supplied by Clifford Frondel and C. S. Hurlbut of Harvard University. E. S. Larsen, Jr., kindly contributed samples of the accessory zircon that he had carefully separated from igneous rocks in Idaho and the San Juan district of Colorado for lead age determinations. These were particularly valuable in providing material of minimum irradiation. He also supplied the sample of pegmatite zircon from Oklahoma which he had succeeded in dating by lead measurements. J. B. Mertie, Jr., of the U.S.-G.S., sent the sample of zircon from the Max Patch granite, from about 3 miles north of Max Patch Mountain, North Carolina, close to the Tennessee line. The accessory zircons from the Nova Scotia granites and the Creighton granite near Sudbury, Ontario, had been previously separated at Massachusetts Institute of Technology for helium studies.

The investigation required the ages of the zircon samples to be known as closely as possible. Since zircon from pegmatites is not too common, in some cases the age had to be inferred from the most tenuous geological relationships. However, the measurements were made on this material in the expectation that lead ages would become available on the same samples at a later date.

TABLE 1.—SPECTROGRAPHIC MEASUREMENTS OF Pb IN ZIRCON

(By M. K. Smith)

Sample No.	Locality	Activity, I α/mg./hr.	Pb* %	Lead age ratio, 10^8 yrs.
1800	N. Hastings Co. Ont.	640	.020	8
1799	N. Hastings Co. Ont.	1380	.035	7
1795	Haliburton Co. Ont.	1220	.0245	5
1804	York R. Quarry, Ont.	245	.008	8
1678	Hammond, N. Y.	510	.013	7
1643	Mineville, N. Y.	180	.005	7
1677	Lewis Co., N. Y.	219	.013	15
1680	Edenville, N. Y.	192	.010	14
1674	Franklin, N. J.	340	.022	16
1683	Franklin, N. J.	187	.008	11
1644	N. Carolina Beach Sands	100	.002	5
1661	Iredell Co., N. C.	212	.002	2
1662	Henderson Co., N. C.	288	.004	4
1669	Buncombe Co., N. C.	615	.005	2

*.002% Pb subtracted from all values to allow for contamination; see text. Precision error on lead measurements less than contamination correction. Error in accuracy unknown for want of absolute zircon standard.

Although Hewitt stated that the Ontario pegmatites are associated with granitic masses of different ages as indicated by crosscutting relationships, a single age of 1000 million years (Nier, 1939) was assigned to all of the samples owing to lack of knowledge on the magnitude of the age differences.

Practically nothing was known of the ages of the group of zircons from the Precambrian areas of New York and New Jersey. This is particularly unfortunate since in this investigation this group controls the upper part of the curve showing the results of the measurements. Rather than assume an age from the very poor existing data, it was decided that rough lead ages (Table 1) obtained from the zircon samples themselves by a spectrographic determination of the lead would probably be closer to the truth, although not to be taken in themselves as individually reliable measurements. The analyses were made by M. K. Smith and G. Shumway, under the direction of L. H. Ahrens at Massachusetts Institute of Technology, and were part of an investigation on improving the accuracy of lead determination in minerals by means of the spectrograph. The samples were arced with 20% NaCl, and .00156% In_2O_3 for an internal standard. On the basis of these results, an age of 1400 million years was assigned to all of the New Jersey and New York samples. The results of the present investigation appear to confirm this approximate age, although the samples could be closer to 1000 million years and still fit a satisfactory curve passing through the points from the Ontario zircons.

Geological evidence seems to favor a late Paleozoic age for at least some of the areas of granite masses and pegmatites in North Carolina. However, much uncertainty remains, and some believe that there are Precambrian intrusives in the region also. Spectrographic lead determinations on four of the samples (Table 1) are very uncertain owing to the small amount of lead present, and to the need of subtracting a fixed amount of lead (.002%) to cover an average value found substquently for contamination from the mortar in the grinding of the samples. An age of 250 million years has been assigned to this group, based on published ratios of total Pb/U in uraninites from this region (Holmes, 1937).*

Except for the Creighton granite, the ages of the accessory zircon samples separated from granites can probably be estimated more reliably. However, since this group is young there

* In a paper received after this manuscript had been submitted, John Rodgers (1952) concludes that the North Carolina pegmatites are probably 350 M.Y. in age on the basis of the published measurements on the Spruce Pine area.

is a question on what the term "age" means in a study of radiation damage. All that can be said at this time is that the age refers to the time since the crystal was cool enough to suffer initial damage by alpha radiation. This may not be much later than the time at which the rock crystallized, as is discussed later in considering the effect of temperature on the process of metamictization.

The Fredriksvarn sample was from the post-Devonian nepheline syenite pegmatite area near the Langesundfjord, Norway, which is believed to be early Permian, and has been assigned an age of 225 M. Y. by Holmes (1937) on the basis of corrected lead ratios. The age of the sample from St. Peter's Dome, Colorado, (860 M. Y.) was obtained by helium ratio (Hurley, 1952)

Method of Analysis

An investigation was made of the most satisfactory parameter that would give a measure of the amount of disorder or degree of metamictization in zircon. An initial attempt to use a densitometer on powder photographs indicated that quantitative work on intensities of diffracted x-ray beams could be done only by means of a spectrometer or other positive linear counting or integrating device. The method of relative solubilities used by Bauer would not indicate inhomogeneity in distribution of the radioelements in the zircon, a knowledge of which must be provided by the method of analysis chosen. The same insufficiency applies to density measurements, but a test was nevertheless made of the density method as a substitute for the x-ray spectrometric method, since the latter equipment is not available to all geological workers. The two samples represented in Table 2 were easily distinguished by 2θ differences, which were in accord with the irradiation data, but not distinguishable by replicate density determinations with the Berman balance. This agrees with Bauer's conclusion that simple density measurements are unsatisfactory.

In the use of the spectrometer, an initial attempt was made to measure the amount of disorder by comparison of peak height (Table 3). The lattice plane (112) was found to combine high intensity diffraction with maximum shift in 2θ and so was used throughout the investigation. A North American Phillips X-ray Spectrometer was used with Cu Kα radiation. The peak height variation between samples did

TABLE 2.—TEST OF DENSITY VARIATION AS A MEANS OF MEASURING RELATIVE DEGREE OF DAMAGE IN ZIRCON

Sample No.	2θ Degrees	Total Irradiation α/mg.	Sample Wt. in air mg.	Density (±.00x)
1659	35.61	23×10^{13}	10.23	4.67
			9.35	4.67
			20.91	4.69
1661	35.58	46×10^{13}	23.10	4.68
			22.28	4.55
			22.94	4.69

show some correlation with total alpha irradiation, but the reproducibility was poor. The powdered samples were packed into slots about half a millimeter deep cut into aluminum plates. The area exposed to the x-rays was maintained constant by the dimensions of the slots, and the sample powder was made flush with the top of the plate to keep the elevation of the surface constant. A number of different methods of sample mounting were tried, but the results were unsatisfactory so that the measurement of peak heights was abandoned.

An attempt to study the area under peaks also was abandoned because, in cases of badly metamict minerals, it was impossible to determine where the base line of the low flat peak should be. The angle representing the center of the peak was then tried, and this was found immediately to be more satisfactory, since there was no longer need for the extremely careful control of the sample area, counting statistics and grain size. It was found that the elevation of the surface of the sample in the spectrometer was quite critical and that diffractions from grains below the surface gave rise to some error. After experimentation it was decided that the very finely pulverized mineral should be smeared onto a glass slide with alcohol to provide a layer one grain thick over an area considerably larger than the x-ray beam. This was found to be highly satisfactory, and the alu-

minum plates were discarded in favor of these glass plate smears.

When a plot of diffraction angle vs. total alpha irradiation was made, a number of points showed a good correlation, but several points were well removed from the group. It was noticed that the x-ray diffraction peaks on these points were invariably skewed, with a sharply rising peak at the high angle end and a tapering lower part of the peak extending toward the low angle end. In these cases the decision as to where to choose the center of the peak became difficult, and it was clear that the material was not homogeneous. Subsequent investigations in which acid was used to leach out the more radioactive parts of the samples showed that this lack of homogeneity was due to an inhomogeneous distribution of the radioactive elements within the zircon crystals.

It eventually became clear that there are three ways in which the radioelements are distributed within the zircon crystals. (1) Zircon has a limited tolerance for uranium and thorium atoms at the time of its crystallization and might take into its structure a uniformly distributed proportion of these elements. This would result in a homogeneous disordering of the material. (2) If conditions changed during the crystallization of the zircon, it might be zoned with a varying proportion of the radioelements in concentric zones in the individual crystals. Then the disordering or metamictization would vary throughout the crystal depending upon the radioelement content of the different zones, so that in a pulverized material some of the grains would be more metamict than others. In this case, there is a gradation in the peak showing a continuous variation from the least metamict zones and highest diffraction angle and peak, to the most metamict zones with lowest diffraction angle and peak. It was found to be impossible to work with these peaks, although an attempt was made to do so by a method that corrected each part of a skewed peak by a multiplying factor that depended on the angle at that point. The factor was found from a plot of peak height vs. diffraction angle on homogeneous samples. However, this attempt was unsuccessful, and it was decided not to use samples showing skewed peaks.

Lastly, it was found that some zircon crystals contain other radioactive crystals as inclusions. In this case, the overall zircon crystal might have a rather uniform level of activity, and a distinctly separate phase representing a spheroidal region about the radioactive inclusions, that is much more highly metamict. The zones about the radioactive inclusions are equivalent to pleochroic halos frequently seen in biotites and represent the volume that is included within the range of the alpha particles coming from the radioactive inclusion. A sample of this type shows a sharp peak representing the uniformly damaged crystal as a whole, plus a low separate peak representing the badly metamict regions about the inclusions at a lower diffraction angle. In these cases, the diffraction angle of the high peak could be measured easily, but the radioactivity of the sample was invariably too high as a result of the contaminating influence of the small radioactive inclusions. One attempt to remove these radioactive inclusions by differential solution resulted in an activity one-fifth of the originally measured activity, causing the point to fall where it should on the plot in Figure 2. However, since these cases of well defined double peaks are rare and selective solution by acid treatment is an undesirable procedure, it was decided that only symmetrical peaks would be used in the correlation. An example of an unsymmetrical peak is shown in Figure 1. As a general rule any material with an activity greater than 500 alphas/mg./hr. should be suspected of having either inclusions of other radioactive minerals or else, particularly in the case of accessory zircons from granites, a highly radioactive skin layer.

As a final measure to improve the precision, powdered silicon was added to the sample to act as an internal standard. The silicon peak at 28.466° (2θ) was measured at the same time and on the same mount as the unknown. The powders were stirred into a mixture of collodion and amyl acetate which formed a uniform coating of the powder on the glass slide. Since the silicon acted as an internal standard, it was no longer necessary to control the elevation of the surface of the powder with the degree of precision needed in the alcohol mounts.

Results of Analyses

Measurements of diffraction angle and activity are listed in Tables 3 and 4, and plotted

FIGURE 1. CONTRAST IN SYMMETRY OF X-RAY PEAK DUE TO INHOMOGENEOUS IRRADIATION OF SAMPLE

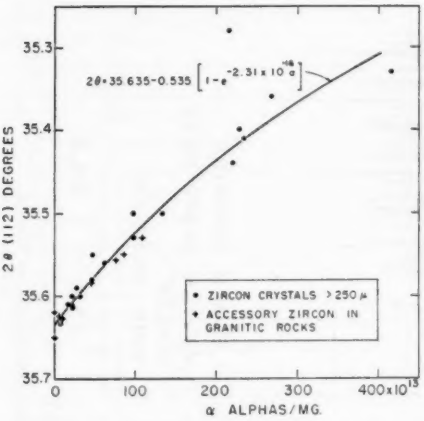

FIGURE 2. DIFFRACTION ANGLE VS. TOTAL ALPHA DISINTEGRATIONS PER MILLIGRAM
CuK α radiation using (112) lattice plane.

in Figure 2. Alpha activity was measured by thick-source alpha count, and by a few thin-source alpha counts (for a check), and in many cases by additional radon measurement to determine the ratio of uranium to thorium.

In the case of pegmatite zircon where the crystals were very much larger than the alpha particle ranges, the "activity index" I, giving alphas/mg./hr., was found as follows. The half-space alpha emission above a thick source of the powdered sample was measured in an ionization chamber. The alpha emission in alphas/cm.^{2}/hr. was reduced to radium equivalent by using a value of μd for zircon of 2.22×10^{-3} gm./cm.^{3} to correct for source absorption, (Nogami and Hurley, 1948), where μ is the ratio of the range of any alpha particle in the mineral to its range in standard air, and d is the density of the mineral. By "radium equivalent" is

TABLE 3.—MEASUREMENTS ON ZIRCON OF GREATER THAN 250 MICRON SIZE

Sample No.	Locality	Relative Peak Height, ins.	2θ(112) Degree	Activity, I Alphas/mg. hr.	Estimated Geologic Age 10^8 yrs.	Total Irradiation 10^{15} α/mg.
1642	Fredriksvarn, Norway	6.0	35.61 ± .0125†	80 ± 12	2.3	16
1643	Mineville, N. Y.	1.9	35.44	180 ± 19	14	220
1644	Beach Sands, N. C.	4.9	35.61	100 ± 11	2.5	22
1659	Renfrew Co., Ont.	7.4	35.61	26 ± 3	10	23
1660	Brundenell Twp., Ont.	6.1	35.59	31 ± 3	10	27
1661	Iredell Co., N. C.	4.1	35.58	212 ± 20	2.5	46
1662	Henderson Co., N. C.	4.0	35.56	288 ± 32	2.5	62
1669	Buncombe Co., N. C.	3.3	35.50	615 ± 68	2.5	134
1670	Lanark Co., Ont.	4.0	35.55	52 ± 6	10	46
1674	Franklin, N. J.	1.8	35.33	340 ± 37	14	415
1677	Lewis Co., N. Y.	0.9	35.36	219 ± 24	14	268
1678	Hammond, N. Y.	2.3	(35.33)*	510 ± 56	14	625
1680	Edenville, N. Y.	3.1	35.41	192 ± 21	14	234
1681	Leeds Co., Ont.	4.4	35.60	36 ± 4	10	32
1683	Franklin, N. J.	4.2	35.40	187 ± 20	14	228
1685	St. Peter's Dome, Colo.	8.1	35.63	11.5 ± 1.2	8.6	8
1794	Glamorgan Twp., Ont.		(35.58)*	510 ± 80	10	450
1794 A	Ditto, treated with HNO_3		(35.53)*	113 ± 12	10	99
1795	Haliburton Co., Ont.		(35.42)*	1220 ± 190	10	1070
1799	N. Hastings Co., Ont.		(35.53)*	1380 ± 220	10	1200
1800	N. Hastings Co., Ont.	0.8	(35.40)*	640 ± 100	10	560
1801	Madawaska, Ont.	0.2	(35.20)*	5750 ± 900	10	5000
1804	York R., Ont.	1.4	35.38	245 ± 27	10	215
1805	Haliburton Co., Ont.	0.2	(35.25)*	1560 ± 250	10	1350
1819	Max Patch, N. C.	5.2	35.60	101 ± 16	2.5	22
1901	Oklahoma pegmatite		35.50	175 ± 28	6.3	97

* Skewed peak; 2θ determined on most ordered phase only; not included in Fig. 2.
† The precision standard deviation of .0125° for 2θ measurements was determined on over 100 repeat measurements on a number of symmetrical peak cases. Errors are greater for skewed cases, depending on amount of skewness.

TABLE 4.—MEASUREMENTS OF ZIRCON WITH SIGNIFICANT WHOLE-CRYSTAL EMISSION
(Mostly accessory zircon from granitic rocks)

Sample No.	Locality	2θ(112) Degrees	Activity, I Alphas/mg. hr.	Whole-Crystal Emission, E Alphas/mg. hr.	Estimated Geologic Age 10^8 yrs.	Total Irradiation 10^{12} α/mg.
305	Creighton Granite, Ont.	35.53	204	110	8	108
1706	Nova Scotia Granite	35.56	400	100	2.5	77
1710	Nova Scotia Granite	35.55	455	125	2.5	86
1887	Hinsdale Rhyolite, Colo.	35.62	393	90	0.03	0.9
1888	Alboroto Intrusive, Colo.	35.625	550	130	0.15	6.4
1889	Alboroto Intrusive, Colo.	35.65	141	33	0.15	1.6
1890	Rubideoux Granite, Calif.	35.58	770	185	1.0	47
1891	S. Calif. Batholith	35.61	(1240)*	290	1.0	(76)
1892	Idaho Batholith	35.61	(2400)*	650	1.0	(145)
1893	Idaho Batholith placer	35.63	170	23	1.0	11.0

*Indication of radioactive xenotime and monazite contamination in samples; measurements rejected.

meant the amount of radium in equilibrium in the uranium series if all of the activity were due to the uranium series, and none to the thorium series. Using the average Th/U ratio mentioned below, the value for I is given as 0.99 times the radium equivalent.

Separate radium analyses were made on a number of samples by fusion of the sample in a vacuum furnace and measurement of the equilibrium radon in an ionization chamber. The results (Table 5) show an average thorium/uranium ratio for the pegmatite zircon of 1.1.

In the case of accessory zircon separated from granites (Table 4), the dimensions of the crystals were small enough so that a significant proportion of the alpha particles originating within the crystals would emerge from the crystals before coming to rest. In these cases, a thick-source alpha count was first made to determine the total number of alphas/mg./hr. generated within the sample. From this figure there was then subtracted a fraction of the count of the alpha particles emerging from a layer of isolated single crystals of known gross weight, this count being designated as E, the whole-crystal emission. The measurement of the whole-crystal emission is made on alpha particles which have a residual air range after leaving the crystal of at least ½ an air-centimeter. Alphas with less than this air range would not cause enough ionization in the gas of the ionization chamber to be recorded. Half of the value on the whole-crystal emission is subtracted from the value of I, on the basis that all of the structural damage caused by alpha particles occurs at the end of their travel, but that approximately half of the total damage per nuclear disintegration results from the recoil of the parent atom, as is discussed later.

TABLE 5.—RADIUM ANALYSES YIELDING THORIUM VALUES BY DIFFERENCE

Sample No.	Alpha Count Radium Equivalent 10^{-12} gm./gm.	Radon 10^{-12} curie/gm.	Calculated Thorium 10^{-6} gm./gm.	Calculated Uranium 10^{-6} gm./gm.	Th/U
1642	83	64	181	180	1.0
1643	185	128	540	360	1.5
1644	101	76	240	215	1.1
1659	27	21	57	60	1.0
1660	31	24	67	68	1.0
1661	215	121	900	340	2.6
1662	293	226	640	640	1.0
1669	615	240	360	680	0.5
1670	53	51	19	144	0.1
1674	348	306	400	860	0.5
1677	224	191	315	540	0.6
1678	520	340	1700	960	1.8
Average					1.1

The measured value of I or (I − E/2) is multiplied by the estimated age of the mineral sample to give a figure for total irradiation, in terms of alphas per milligram. These values are plotted against the values of the x-ray diffraction angle, 2θ, from the plane (112) (Fig. 2). The sine of this x-ray diffraction angle is inversely proportional to a dimension of the unit cell of the zircon, so that smaller angles repre-

sent a distention of the structure due to disordering effects.

Expression for Metamictization

In Figure 3 is shown a plot of relative peak height on the x-ray spectrometer record versus

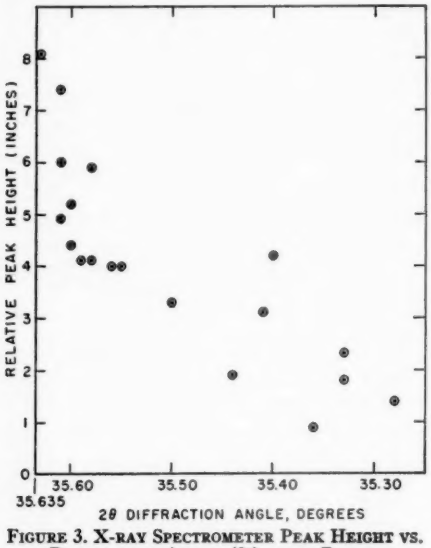

FIGURE 3. X-RAY SPECTROMETER PEAK HEIGHT VS. DIFFRACTION ANGLE (2θ) FROM ZIRCON PLANE (112)

the x-ray diffraction angle, 2θ, at the center of the peak. The scatter of the points is due to the difficulty in preparing mounts of constant grain density and area, and is the reason why this method of measuring the amount of disorder was discarded. However, the plot shows that, as the mineral sample becomes more metamict, the peak height decreases and the diffraction angle becomes smaller; and in the cases of the most metamict zircons the angle approaches 35.2°. Actually, in these highly metamict zircons, the least disordered phase shows the strongest peak so that there is a tendency for the angle to be read at a point somewhat above the true angle representing the most metamict phase of the sample. In these cases the peaks are low and flat, tapering off to the base line well below the angle 35.2°. It has therefore been estimated that an angle of 35.1° more closely represents the diffraction angle for completely metamict material.

At the least disordered end of the scale, Figure 2 shows that an extrapolation of the curve to zero alphas per milligram gives an angle of approximately 35.635° for the value of 2θ for completely undamaged zircon. Furthermore it seems from the plot of points that the process of metamictization or crystal structure damage is an exponential one, leveling off at some value for 2θ beyond which no further damage can be done.

An expression for this process can be conceived as follows. It might be assumed that zircon becomes progressively disordered simply because of a very slow annealing rate compared to such minerals as sphene and monazite, which do not become metamict to any extent despite comparable alpha activity.

Let N = number of displaced atoms per mg.
N_ϵ = limiting value of N at maximum disorder
λ_M = fraction of displacements returning to correct structure position per million years at minimum temperature (earth's surface temperature)
K_1 = number of atoms displaced per alpha disintegration
A = number of alpha disintegrations/mg./M.Y. (assumed constant as sufficient approximation)

For any single homogeneous volume of crystal

$$\frac{dN}{dt} = AK_1\left[\frac{N_\epsilon - N}{N_\epsilon}\right] - \lambda_M N \qquad (1)$$

$$\ln\frac{\left[AK_1 - N\left(\frac{AK_1}{N_\epsilon} + \lambda_M\right)\right]}{C_1} = -\left(\frac{AK_1}{N_\epsilon} + \lambda_M\right)t$$

$$AK_1 - N\left(\frac{AK_1}{N_\epsilon} + \lambda_M\right) = C_1\, e^{-((AK_1/N_\epsilon)+\lambda_M)t}$$

$$\text{At } t = 0, \quad N = 0; \quad C_1 = AK_1$$

$$\frac{N}{N_\epsilon} = \frac{AK_1}{(AK_1 + N_\epsilon\lambda_M)}\left[1 - e^{-((AK_1/N_\epsilon)+\lambda_M)t}\right] \qquad (2)$$

If N/N_ϵ is approximated by $(\theta_i - \theta/\theta_i - \theta_\epsilon)$ where θ is the doubled diffraction angle for the (112) plane, and θ_i, θ_ϵ are the same angles for zero and fully metamict crystals respectively, then:

$$\frac{\theta_i - \theta}{\theta_i - \theta_\epsilon} = \frac{AK_1}{AK_1 + N_\epsilon\lambda_M}\left[1 - e^{-((AK_1/N_\epsilon)+\lambda_M)t}\right] \qquad (3)$$

From Figures 2 and 3, θ_t may be taken as 35.10° and θ_i as 35.635°.

Let $\Delta = 35.635 - \theta$ and $\alpha =$ At alphas/mg. If λ_M is small enough to be insignificant,

$$\Delta = 0.535\,(1 - e^{-(K_1 \alpha / N_t)}) \qquad (4)$$

and the crystal structure decays exponentially into a metamict state simply as a function of α, or the total irradiation, regardless of time.

If, on the other hand, λ_M is significantly large compared to (AK_1/N_t), the disorder is a function of both α and t, and each zircon sample would fall on a different curve depending on its activity. Such a family of curves is shown in Figure 4, since this is probably the case for minerals such as sphene and monazite which reach an equilibrium at low levels of disorder.

In the case of zircon it would appear that the annealing rate, λ_M, is not significant compared to the breakdown rate (AK_1/N_t), since young zircons of high activity fall on the same curve as old zircons of low activity. This equation (4) is therefore used for a theoretical curve to fit the points in Figure 2. The best value for "half-metamictization" was chosen at an irradiation of 300×10^{13} alphas/mg. This gives a value for (K_1/N_t) of 2.31×10^{-16}.

If N_t amounts to essentially all of the atoms in a mg. of zircon, namely, 2×10^{19}, then this particular curve gives a value for $K_1 = 4.5 \times 10^3$ atoms displaced per alpha particle.

This value for K_1 is illuminating. It shows in the first place that the disordering process is not due to helium atoms acting as impurities. It further indicates that the disordering is not caused by rare events, but that most of the possible displacements per alpha by atomic collision (see Discussion below) have occurred, and that the displaced atoms remain permanently out of lattice position. This may be because the recrystallization and reorientation observed at high annealing temperatures is too sluggish at earth's surface temperature to be observable, or because of the formation of a more stable form for the components at the low temperatures. If baddeleyite (ZrO_2) is formed, it does not form large enough crystals to give a diffraction pattern, and the same applies to SiO_2. The appearance of a baddeleyite diffraction pattern at intermediate annealing temperatures suggests that this structure may be the more stable low temperature form.

The disordering process is therefore visualized as an increasing number of small localized aggregates each of which contains 10^3 to 10^4 disordered atoms (probably as an aggregate of ZrO_2 and SiO_2) caused by the individual alpha particles and the recoiled parent atom. Each

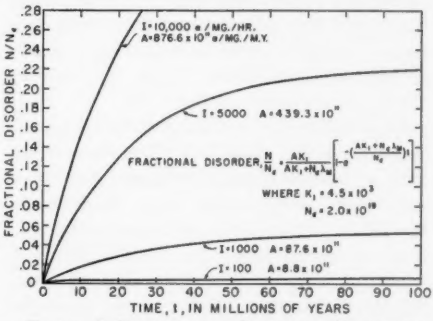
FRACTIONAL DISORDER N/N_t
TIME, t, IN MILLIONS OF YEARS
I=10,000 α/MG./HR.
A=876.6 x 10^10 α/MG./M.Y.
I=5000 A=439.3 x 10^10
I=1000 A=87.6 x 10^10
I=100 A=8.8 x 10^10
FRACTIONAL DISORDER, N/N_t = AK_1/(AK_1+N_tλ_M) [1 - e^-((AK_1+N_tλ_M)/N_t)t]
WHERE K_1 = 4.5 x 10^3
N_t = 2.0 x 10^19

FIGURE 4. EXAMPLES OF PREDICTED RELATIONSHIP BETWEEN DISORDER AND TIME, FOR MINERALS WHICH RECRYSTALLIZE AT EARTH'S SURFACE TEMPERATURE

E.g. recrystallizing (annealing) rate, $\lambda_M = .0347$/M.Y. or equivalent recrystallizing half-life = 20 M.Y.

alpha event probably causes two of these disordered volumes; the parent atom causing the volume at one end, and the last 40,000 ev. of the alpha particle's energy causing the volume at the other end. This is a change of view from that stated in a previous paper (Hurley, 1952) in which the ionization caused by the alpha particle in its loss of energy down to nearly zero was believed to be responsible for the major part of the disorder.

Effect of Temperature of Environment

Tests were made on the annealing of partially metamict zircons at elevated temperatures. The results confirmed those described in the literature (Pabst, 1952), and it was found that if a crystal were not greatly disordered it would return to a degree of order very close to that of a perfect crystal, namely, with a diffraction angle close to 35.63°. The fact that λ_M, the annealing rate for zircon at earth's temperature, has been found to be essentially zero is in contrast with the relatively rapid annealing rates for most other minerals at this temperature. Recent x-ray investigations by the writers on sphene, apatite, and quartz indicated that these minerals show little or no damage despite irradiations comparable to those of the zircon samples. Other cases mentioned in the Introduction

are well known. The fact that zircon inclusions in feldspar and other common rock minerals do not show isotropic haloes about the zircon grains is proof that these minerals also reach an equilibrium between damage rate and recovery that is at a low level of damage, as in one of the lower curves in Figure 4. Glasses formed in effusive rocks crystallize rapidly, and are seldom found in the vitreous state earlier than the Tertiary.

Thus it seems probable that there is a transition temperature above which the zircon structure is the most stable and below which some other form (probably baddeleyite) is most stable, other factors remaining constant. There is some indication from these results that this transition temperature is fairly high, since the samples of zircon separated from young granites fit the curve in Figure 2 on the basis of an estimated age for the granite, despite the fact that these granite masses must have taken several tens of millions of years to rise and cool.

This inference is in agreement with the x-ray findings mentioned previously in which the baddeleyite structure appeared at temperatures near 900°C. Thus this transition temperature is probably about 500°C. under high pressure aqueous conditions (from the geological occurrence of zircon) and higher under earth's surface conditions. This would mean that a small decrease in temperature and pressure below the conditions under which the zircon was formed would bring the mineral into the range in which disordered material would not return to the zircon structure. Thus instead of a decreasing annealing rate that is temperature-dependent over a significant part of the life of the mineral, it appears likely that the annealing rate is zero after a small drop in temperature and pressure, and that the time measured by this method would be not many millions of years less than the age since the zircon crystallized.

Information of Interest in the Study of Radioactive Heat Generation in Rocks

The question has been raised many times whether a large proportion of the energy given off in radioactive disintegrations in rocks is stored in the rocks in the form of crystal structure dislocations, or whether most of this energy is released immediately as heat. The results of this investigation answer this question conclusively. In the case of zircon, it can be seen that the total energy required to make the structure half metamict is equivalent to 3×10^{15} alphas per milligram. The average alpha energy in the uranium and actino-uranium series in 5.39 mev. per alpha, and in the thorium series 6.03 mev. per alpha. Adding the energy due to betas and gammas and allowing 70% of the activity to be due to the uranium series, the mean energy per alpha is 5.99 mev. One mev. is equivalent to 0.382×10^{-13} calorie. Therefore 685,000 calories per gram have been released in the zircon during the process of half-metamictization. Since it only requires one two-thousandth of this amount of energy to melt the zircon structure, it is clear that a maximum of only about .02% of the energy can be retained in the structure in the form of disordered atoms, or even much less energy if the disordered atoms have recrystallized into a fine aggregate of some more stable form such as baddeleyite. In the case of other minerals the amount of energy retained is negligible since most of these minerals do not show any effect of radiation. Thus it is clear that in the question of radioactive heat production in rocks, essentially all of the energy of radioactive disintegrations is released immediately as heat within the rock.

The question of how much energy can be stored within the rock over a long period of time in the form of atomic disarrangement depends on a knowledge of the equilibrium disorder in feldspar, quartz, and the ferromagnesian minerals. Since only the host minerals showing isotropic haloes[2] and the few source minerals that become metamict are capable of storing any appreciable energy it is evident that the maximum amount of energy stored at equilibrium in any normal rock type will be of no significance in questions of petrogenesis and volcanism.

Information of Possible Use in Deriving Uranium from Zircon

In many granitic rocks about half of the uranium is tied up in the accessory zircons. Most

[2] An interesting example of a rock-forming mineral that breaks down under alpha radiation is cordierite. Excellent illustrations of haloes in this mineral are given by Mügge (1927, Plate 4).

ites contain 10^{-3}–10^{-4}% of uranium and three times as much thorium, and the accessory zircons measured in this investigation contain about 0.1% of uranium and 0.1% of thorium. Special cases may be higher by an order of magnitude. Zircons may be obtained fairly easily from granitic rocks within a few hundred feet of the surface, where weathering has loosened up the grains, by crushing the rock to pass 20-mesh and using gravity methods of separating the heavy constituents. Zircon is fairly easily separated from other heavy constituents under strong magnetic fields. A more accessible source of zircons is that of the natural concentrations of this mineral in certain beach sands.

Where not strongly metamict, zircon is a particularly refractory mineral, both for acid leaching and fusion. It is fortunate that in many zircons, and particularly in the radioactive ones, the radioactivity coincides with the most metamict part of the zircon, and this part is most susceptible to acid leaching. Thus a simple acid leach by hydrofluoric or even nitric acid will remove most of the uranium from the zircons, if they are ground finely enough. Since metamictization is a function of both activity and age, the quest should be for zircons not only active but fairly old, and from syenitic rocks rather than granitic. As a first approximation, the radioactivity of the zircons and the possibility of a good recovery by acid leach may be forecast quite simply by x-ray diffraction analysis.

Discussion of Process of Radiation Damage

The effects of radiations and ionizing particles on materials of various kinds are the subject of considerable study at the present time owing to the increasing need for shielding and structural materials suitable for applications of nuclear science. The study of radiation damage in geological materials of great age is therefore of interest, since it is possible to have equilibria established between damaging events and annealing events at temperatures much lower than possible for the short times available in artificial radiation studies. Slater (1951) has summarized the theoretical considerations involved in radiation and particle damage, giving particular weight to the effect of atomic displacements and of impurities resulting from the particles coming to rest within the solid. An alpha particle passing through an insulator such as zircon dissipates most of its energy by exciting the electronic system, producing free electrons and excitons. (For discussion, *see* Seitz, 1952.) These excitations occur primarily along the high energy part of the alpha path during the time that the alpha particle has a positive charge of at least one. It is doubtful whether these electronic excitations will displace atoms, producing interstitial atoms and vacant lattice sites. Electrons will be removed as long as their orbital velocities are less than the velocity of the particle, so that an alpha particle continues to ionize its host until its kinetic energy drops to about 40,000 electron volts, after which time most of the remaining energy is lost by atomic collision. Therefore, for a 6 mev. alpha particle, roughly 99 per cent of its energy is dissipated through excitation of the electronic system and less than 1 per cent through atomic collisions, the latter occurring primarily at the end of the travel. However, this 1 per cent or less of the alpha particle energy, together with the recoil of the parent atom, may account for a large part of the radiation damage to the crystal.

Since the atoms are held in their lattice positions by energies in the neighborhood of 25 ev, each alpha particle losing a few tens of kilovolts by atomic collision would therefore be expected to free not more than 10^3 atoms. The kinetic energy passed on to atoms not only in primary encounters, but also in the secondary and tertiary ones, and in fact all those with energies of 25 volts and up, is so great as to correspond to an exceedingly high temperature, of the order of several hundred thousand degrees Centigrade. Even when the average energy per knock-on atom is reduced to one volt, by interchange with its neighbors, so that 10^5 such atoms are involved, the energy of each atom still corresponds to more than the boiling point, even for a refractory material. According to Slater, there will be a track of high-temperature vaporized material about the end part of the track of the particle. Following this, there would be an extremely rapid quenching, resulting in a disordered aggregate. If this disordered

region crystallizes again, orientation may be controlled by the ambient lattice surrounding it, so that it will grow again into the crystallographic orientation of the host mineral. Thus it is expected that there will be a relatively small proportion of isolated displaced atoms and vacant lattice sites along the major path of the alpha track, but that most of the displacements will be in a completely disordered volume at the end of the track.

A similar statement may be made regarding the recoiled parent atom. This atom will be recoiled with an energy of approximately a sixtieth of that of the alpha particle, or about 10^5 electron volts. This energy would be largely dissipated by atomic collisions in the same way as the alpha particle at the end of its path. Since the removal of electrons by a travelling particle is dependent upon its velocity rather than its energy, the recoiled parent atom would fail to displace electrons at a considerably higher energy than would an alpha particle. Slater estimates this transition energy for a proton to be less than 10,000 electron volts, for an atom of intermediate weight (for instance, a carbon atom) about 100,000 electron volts, and for a heavy atom, such as the recoiled parent atom, to be over a million electron volts. Thus the number of atoms displaced by both the alpha particle and the recoiled parent atom may be as many as 5000. This agrees with the value found for K_1 in this investigation.

In considering the question of structural disorder produced by foreign atoms introduced into the mineral, in this case helium and lead atoms, it has been found that only one foreign atom per 1000 atoms in the structure has been introduced by the time the mineral is metamict. This would not appear to be a sufficient cause for the complete disordering of the structure, and is therefore excluded as a possible mechanism.

There has been a natural tendency to associate structural damage with electronic excitation produced by an alpha particle, and to think of the damage occurring along the length of the track. This is because the particles have always been observed and measured by the ionization they produce. However the blackening along alpha tracks in nuclear track emulsions is a chemical effect triggered by the ionization, and the condensation in a cloud chamber is a similarly triggered physical effect. Also the thermoluminescence in minerals, although due to the energy of trapped electrons, involves small amounts of energy compared to that involved in displaced atoms. Since the magnitude of K_1 from the results presented herein is in accord with the amount of alpha particle energy dissipated by atomic collisions, it is concluded that electronic excitations do not appreciably damage the structure. The primary accelerated electrons have a considerable range and the energy imparted to the electronic system is spread into a considerable volume, dissipating the heat without atomic displacements. Of particular value as evidence in favor of this conclusion is that metamict biotite occurs only as shells about a radioactive inclusion, as seen in the best-defined examples of pleochroic haloes, and not in the region enclosed by the shells.

Metamictization of Uraninite

So far the discussion has been focussed on the primary effects of radiations. Secondary effects may occur as a result of chemical reactions with materials of the environment.

Natural uraninite when fresh and unweathered usually shows the same sharp diffraction pattern as freshly prepared synthetic uraninite. This indicates that disordered atoms return to the isometric structure of UO_2 so readily that the equilibrium between alpha damage rate and recrystallization is at a low level of disorder. Thus the metamictization of uraninite when it does occur is evidently not due to the primary disturbance of atoms by atomic collisions as in the case of zircon.

Results of work on the chemistry of uranium oxides reported by Katz and Rabinowitch (1951) are illuminating. These authors state that UO_2 represents a solid solution series extending from possibly as low as $UO_{1.75}$ (which may be a separate homogeneous phase, U_4O_7) to $UO_{2.30}$. This has the fluorite structure. Above $UO_{2.30}$ there follows a diphasic region, the upper limit of which is not definitely established. Whether the next monophasic range begins at $UO_{2.50}$ or at $UO_{2.62}$ is uncertain, but there is no doubt that the upper limit of this homogeneous range extends far beyond $UO_{2.67}$—in all probability up to UO_3. The phases between $UO_{2.6}$

(or $UO_{2.62}$) and $UO_{3.00}$ show a continuous transition from an orthorhombic to a hexagonal lattice. In addition to the hexagonal form, uranium trioxide also exists in one amorphous and in at least two, and perhaps four, other (as yet unidentified) crystalline forms.

Thus it seems likely that, in the process of metamictization of uraninite, although no change of structure takes place due to alpha radiation itself, the large number of accelerated electrons changes the oxidation state of some of the uranium, and the structure becomes ready to accept more oxygen if such is available from the outside. If more oxygen is available, the displaced atoms caused by atomic collisions will then not anneal back to the UO_2 structure, but return to the more stable orthorhombic-hexagonal range, and the mineral will become metamict by oxidation. This is supported by results reported by Brooker and Nuffield (1952) who measured cell dimensions, intensity of x-ray diffraction, and ratio of U^6/U^4 in pitchblende samples from the same deposit, and found that these quantities were related to the amount of weathering that each sample had suffered. The reports of Ellsworth, Tomkeieff, and others note alterations that vary from the surface to the interior of uraninite crystals, or crystals that have become zoned or locally metamict.

Thus it appears that uraninite may become metamict by a chemical alteration dependent on its environment, as opposed to zircon which breaks down continuously as a function only of total irradiation.

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PETROLOGY OF GRANOPHYRE IN DIABASE NEAR DILLSBURG, PENNSYLVANIA

BY PRESTON E. HOTZ

ABSTRACT

Small bodies of granophyre occur in the upper part of diabase bodies of Triassic age in southeastern Pennsylvania. One near Harrisburg was penetrated by a diamond-drill. Drill core specimens show a gradation from diabase to granophyre.

New data include 10 chemical analyses, spectrographic determinations of trace elements, and the results of petrographic study of specimens from the drill core. The sequence, from diabase to granophyre, includes a chilled zone that represents an original magma of tholeiitic composition, normal diabase, pegmatitic facies of diabase, and granophyric diabase that is intermediate in composition and petrographic characteristics between diabase and granophyre, and finally granophyre. Alkalies and silica increase progressively from diabase to granophyre; iron increases to a maximum in transitional granophyric diabase, then decreases in the granophyre.

It is concluded that crystal fractionation in a large sheetlike body of tholeiitic magma yielded a small amount of granophyre. Prior to complete solidification, a residual liquid rich in iron, alkalies, and silica accumulated locally in the upper part of the diabase sheet. In places volatile-rich iron-bearing solutions escaped into the overlying sedimentary rocks and deposited magnetite; the remaining liquid crystallized to fine-grained granophyre.

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Introduction

Terrestrial and lacustrine sedimentary rocks of Triassic age occupy several troughlike basins occurring with some interruptions from Nova Scotia to South Carolina. The basins are narrow and are elongate north and south; and range from 25 to more than 300 miles in length and from 10 to 30 miles in width. The largest extends from southeastern New York to northern Maryland; its northern part is commonly known as the Newark Basin and the southern part as the Gettysburg Basin. Toward the end of the Triassic period, basaltic magma invaded these basins and formed intrusive sheets and flows.

In the Gettysburg Basin and southern part of the Newark Basin are many intrusions of diabase (Fig. 1). Many have a ring-shaped outcrop pattern due to their warped sheetlike habit as described elsewhere (Hotz, 1952). Magnetite deposits are associated with bodies of diabase in southeastern Pennsylvania, the most important and best known of which is at Cornwall, Lebanon County.

The undifferentiated magma that invaded these sedimentary basins represents a worldwide type, the tholeiitic magma type of Kennedy (1933, p. 240–242). The thick sill that forms the Palisades in New Jersey and New York cooled slowly enough to permit a considerable degree of differentiation. Lewis (1908, p. 155–162) and Walker (1940, p. 1059–1105) have studied the petrography, and have described the differentiation process that formed the well-known olivine-rich layer near the bottom of the sill. Stose and Lewis (1916, p. 623–643) described the diabase in the vicinity of Gettysburg, and Stose and Jonas (1939, p. 126–130) described the Triassic intrusions in York County, where they recognized small isolated masses of "pink diabase or diabase pegmatite" as differentiates of the diabase magma.

Dillsburg is a small village in northeastern York County, Pennsylvania, about 15 miles southwest of Harrisburg (Fig. 1). From about 1855 to the early 1900's, a small amount of magnetite was produced in the district from underground workings and shallow open cuts. In 1945, during further exploration of the deposits, drilling located a body of diabase beneath a cover of sedimentary rocks. The upper part of this body contains a zone of pink granophyre grading above and below into apparently normal diabase and affords an unequalled opportunity to make a detailed petrographic study.

Acknowledgments

The field work and part of the laboratory studies for this report were carried on by the writer in 1945 and 1946 as a phase of the study of the Triassic magnetite deposits of the Dillsburg district by the U. S. Geological Survey (Hotz, 1950). This was supplemented by additional independent laboratory study during 1947 and 1948. I am indebted to Princeton University for office facilities and for a research grant to cover the cost of 10 chemical analyses, and to Professors A. F. Buddington and H. H. Hess for their interest and helpful suggestions throughout the course of the investigation.

General Geology

The areal geology of the Dillsburg district has been more or less completely described. The magnetite mines were first described by Frazer (1877). A later detailed report was made by Spencer (1908, p. 74–96). An important discussion of the origin and somewhat different interpretation of the structure of the ore deposits was made by Harder (1910). A map and report on the geology of York County, including the Dillsburg district, was prepared by Stose and Jonas (1939). Two recent articles on the results of diamond drilling exploration of the magnetite deposits have been prepared by Neumann (1947) and the writer (Hotz, 1950).

The oldest rocks in the area are early Paleozoic and occur northwest of Dillsburg, where they are in fault contact with Triassic rocks (Fig. 2).

The Triassic sedimentary rocks are predominantly red shale and sandstone of the Gettysburg shale. This formation is characterized by lenticular beds of limestone conglomerate whose constituent pebbles have been derived from the near-by Paleozoic rocks. The magnetite deposits are replacement bodies in the limestone conglomerate. Within the formation is also a belt of coarse arkosic sandstone and quartzose fanglomerate, which has been mapped

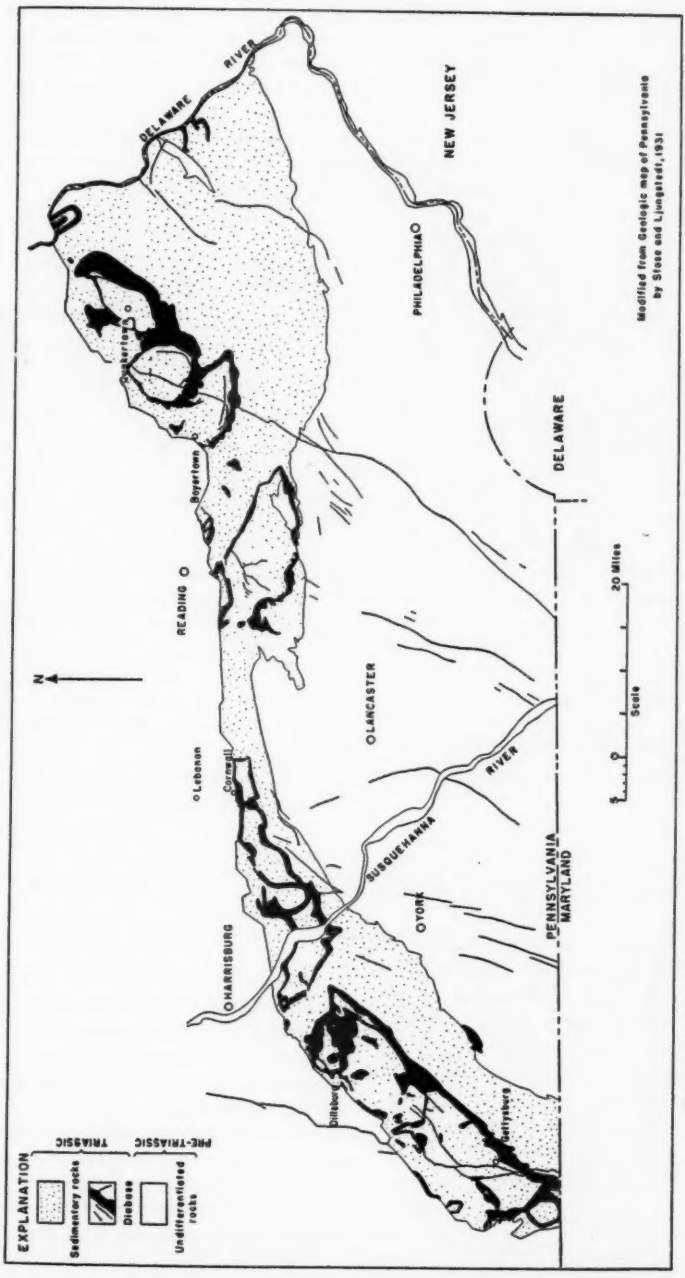

FIGURE 1. MAP SHOWING THE DISTRIBUTION OF DIABASE IN TRIASSIC ROCKS OF SOUTHEASTERN PENNSYLVANIA

and described by Stose and Jonas as the Heidlersburg member of the Gettysburg shale. Adjacent to the bodies of diabase, the normally red sedimentary rocks are bleached and metamorphosed to gray and buff hornfels and quartzite, and the limestone conglomerate is recrystallized to marble or replaced by silicates.

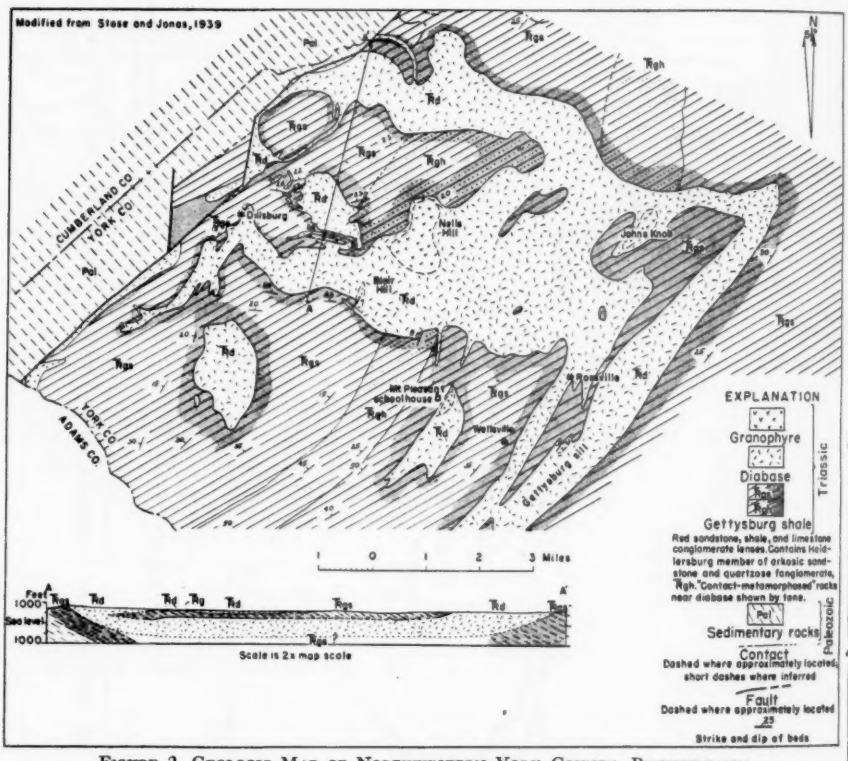

FIGURE 2. GEOLOGIC MAP OF NORTHWESTERN YORK COUNTY, PENNSYLVANIA

In the Dillsburg region (Fig. 2), the main body of diabase is continuous to the south with the long, apparently conformable Gettysburg sill (Stose and Bascom, 1929, p. 11). Westward, this large intrusive mass forms a ring that almost completely encloses the area of sedimentary rocks near Dillsburg.

PETROGRAPHY

Definition of Diabase

To avoid confusion it is desirable at the outset to clarify the use of the term diabase. As used here, diabase includes hypabyssal rock of basaltic composition and ophitic or subophitic texture, which is commonly called dolerite by writers outside North America.

General Characteristics

Diabase.—The diabase of the Triassic basins of the eastern United States is exceptionally uniform. It is commonly gray, medium-grained and of uniform texture. Fresh white feldspar laths intergrown with dark grayish-green to black pyroxene, and scattered black metallic granules of ilmenite magnetite can be identified megascopically in most specimens. The rock of the more quickly cooled smaller intrusions or the borders of large bodies is correspondingly darker, finer-grained and more dense.

Pegmatitic facies.—Bodies of very coarse-grained diabase as much as several feet thick found in the upper parts of the larger diabase

masses are, because of their appearance, occurrence, and genesis, commonly called pegmatitic diabase. Walker (1940, p. 1065) writes of them as "pegmatite schlieren," and (1949, p. 663) "dolerite pegmatites." Shannon (1924, p. 14–22) applied the term "diabase pegmatite" to similar rocks at Goose Creek, Va. Cornwall (1951, p. 163) also described "pegmatitic facies" in basaltic lavas of the Keweenawan series.

Characteristically, the rock is coarse-grained, and the pyroxene crystals, partly altered to amphibole, are long and blade-like, attaining lengths of as much as 3 inches. Between the pyroxene blades is chalky white plagioclase in which there may be some pinkish patches of potash feldspar and a little visible quartz. Skeletal crystals of ilmenite-magnetite are abundant.

Granophyre.—Granophyre is known at several localities in York County. At least six bodies are known in the Gettysburg sill and its offshoots near Dillsburg (Fig. 2). Most of these are shown on the geologic map of York County (Stose and Jonas, 1939, pl. I).

All the bodies of granophyre are small in relation to the size of the diabase bodies enclosing them. They lie in the upper part of intrusive sheets or sills and some are beneath a cover of normal diabase. The bodies are lenticular with their longer axes oriented more or less parallel to the length of the enclosing diabase body.

In contrast to the gray diabase, the granophyre is pale pink. Grain size ranges from fine to coarse. The coarsest-grained granophyre seen in the region is exposed near Mt. Pleasant School.

Miarolitic structures were observed at some localities. They range from microscopic dimensions to almost an inch in diameter. The coarser-grained rocks possess the largest cavities. Stubby terminated prisms of quartz and crystals of pink feldspar project into the cavities, as well as a few octahedra of magnetite. Fibrous masses and needles of green actinolitic amphibole extend into, across, and may partly fill some of the openings. Botryoidal opal encrusts the walls of some cavities in the granophyre at Mt. Pleasant School.

The essential minerals are turbid sodic plagioclase and an abundant micropegmatitic intergrowth of interstitial quartz and alkalic feldspar. The rock may contain some individual quartz grains. Hornblende, rather than pyroxene, is the mafic mineral in most of the granophyres, and is primary rather than a replacement of pyroxene. Where pyroxene is the principal mafic mineral, it is a green hedenbergitic variety. Other accessory minerals include biotite, chlorite and actinolitic amphibole. Ilmenite-magnetite and apatite are common minor accessories.

On the York County geologic map (Stose and Jonas, 1939), a large area of granophyre is shown west and southwest of Nells Hill. Actually only the northwestern third of this area is true granophyre. South of this smaller area the rock is more of a transitional type between diabase and granophyre and contains considerable alkalic feldspar and quartz but also some moderately calcic plagioclase. The rock grades abruptly downward into normal diabase.

The granophyre on Blair Hill near the southwestern boundary of the diabase (see Fig. 2) is a medium-grained, pinkish- to brownish-red rock devoid of miarolitic cavities. Its western and southern contacts are mapped with relative certainty, but the eastern limit is more difficult to locate, partly because of inadequate exposures but mostly because of an apparent transition downward into normal diabase. The unquestioned granophyre contains abundant turbid alkalic feldspar and quartz in large individuals as well as in micrographic intergrowth with feldspar. Green hornblende is the mafic mineral but is not very abundant. The rock judged in the field to be transitional between granophyre and diabase contains abundant interstitial micropegmatite and large euhedral laths of altered plagioclase approximating medium andesine in composition. Hornblende is the essential mafic mineral here also and late-stage fibous green amphibole is plentiful. The rock appears to have been considerably affected by hydrothermal solutions, and possibly the hornblende is secondary after pyroxene, though no pyroxene relicts were observed.

The areas of granophyre east of Wellsville and at John's Knob (Fig. 2) and south of the map area have been visited and typical specimens collected. Some thin sections of these

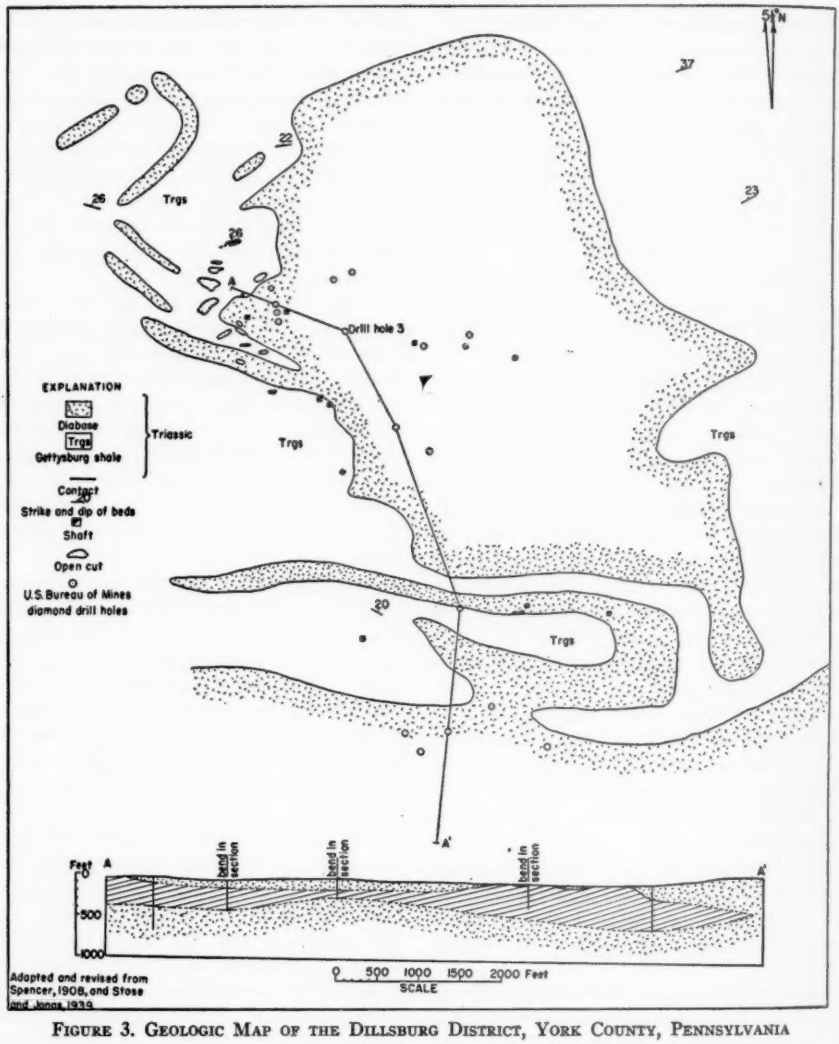

FIGURE 3. GEOLOGIC MAP OF THE DILLSBURG DISTRICT, YORK COUNTY, PENNSYLVANIA

specimens were examined but no detailed petrographic studies have been made of them. Stose and Jonas (1939, p. 127–130) have presented an adequate description of these rocks. All are composed of sericitized alkalic feldspar and quartz as individual crystals and in micrographic intergrowth with the potassic feldspar. Green hornblende is the important mafic mineral and the usual accessories, apatite and ilmenite-magnetite, are present. Allanite is reported by Stose and Jonas (1939, p. 129) in the rock from the John's Knob area and epidote from another.

Diabase-Granophyre Sequence Exposed by Diamond Drilling

During exploration of the magnetite deposits at Dillsburg, a body of diabase beneath the sedimentary rocks in which the magnetite

deposits occur was revealed by diamond drilling (Hotz, 1950, p. 12–13). The diabase body, whose base was not reached by the drilling, appears to lie beneath the whole area in which the magnetite deposits are found, and is believed to be a broadly concave, platter-like, discordant sheet, which is probably an offshoot from the diabase sill that extends north of Gettysburg, Pennsylvania. The thickness of the body probably exceeds 1000 feet. The sheet-like body at Dillsburg and the diabase bodies of Triassic age in southeastern Pennsylvania have been described elsewhere (Hotz, 1952).

One drill hole (No. 3, Fig. 3) penetrated the diabase for 370 feet below its upper contact, and at a depth of about 210 feet penetrated a zone of granophyre more than 100 feet thick. The core from this drill hole made available for study a complete series of specimens from the diabase to granophyre. The drill core was studied and logged in detail and thin sections were cut from specimens taken at intervals of 5 feet; in addition, chemical analyses were made of eight specimens representing the main rock types. The main units or rock types are summarized schematically in Figure 4.

The top of the diabase in contact with the overlying metamorphosed sedimentary rocks of the Gettysburg shale is very fine-grained, but not dense, at the contact. The chilled zone of fine-grained material is about 27 feet thick, and the grain size gradually increases with depth. Within the lower part of the chilled zone and continuing to about 58 feet below the upper contact, the texture is somewhat heterogeneous because of an irregular mesh-work of coarser material in the medium-grained diabase. With increasing depth, the coarser meshwork fades out and the diabase is dominantly and uniformly medium-grained. Sparse local layers of coarser-grained diabase half an inch to 2 inches thick alternate with the medium-grained diabase. In general, the diabase become coarser-grained with depth. A pegmatitic facies of the diabase 10 feet thick was cut by the drill about 180 feet below the roof.

Below the pegmatitic facies the diabase is slightly pinkish. The change from diabase to granophyre is transitional. The pink color becomes more pronounced; there is a noticeable decrease in the content of dark minerals; and the texture becomes somewhat finer-grained.

The upper limit of true granophyre is about 210 feet below the contact of the diabase and

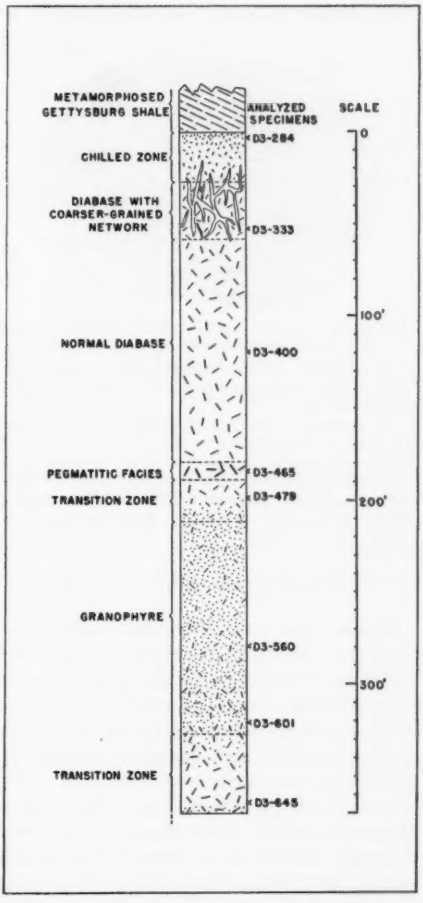

FIGURE 4. SCHEMATIC REPRESENTATIONS OF DRILL HOLE 3

the sedimentary rocks. The lower boundary is also transitional, and the lower limit of the granophyre was placed 318 feet below the top of the diabase; the granophyre zone is about 108 feet thick.

Below a depth of approximately 300 feet, the grain size increases somewhat, dark minerals are once more abundant, and the rock begins

to lose its pink color. More diabasic-appearing rock alternates with rock that is closer to granophyre in appearance. The drill hole was terminated in the transition zone below the granophyre, and therefore normal diabase was not encountered beneath the granophyre.

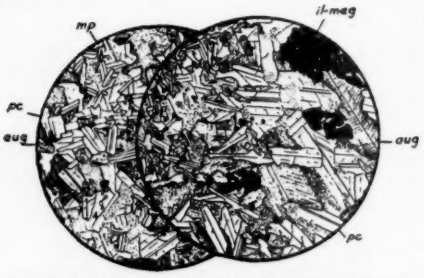

FIGURE 5. DIABASE and FINE-GRAINED DIABASE WITH COARSE-GRAINED NETWORK

a. Diabase. 128 feet below upper contact, drill hole 3 (pc, plagioclase; mp, micropegmatite; aug, augite) × 10.

b. Pegmatitic area in fine-grained diabase. Note absence of well-defined boundary between diabase (left) and pegmatitic facies (right). From zone of fine-grained diabase with coarse-grained network, 46 feet below upper contact, drill hole 3. (pc, plagioclase; aug, augite; il-mag, ilmenite-magnetite) × 10.

Petrography of the Drill Hole Section

General statement.—The essential constituents are plagioclase, micropegmatite, and pyroxene. Accessory minerals include ilmenite-magnetite, hornblende, chlorite, and ilvaite.[1] Minor accessories are apatite, biotite, pyrite and chalcopyrite, epidote, sphene, and calcite.

Chilled contact facies.—The rock throughout the chill zone is holocrystalline. The average grain size ranges from 0.1 mm. to 0.5 mm. Colorless to faintly greenish augite is intergrown with laths of plagioclase (An_{65-70}) in typical diabasic or subophitic texture.[2]

Tiny angular to subangular grains of ilmenite-magnetite are scattered uniformly through the rock and included in the plagioclase and pyroxene. A small amount of quartz occupies the interstices between earlier crystallized constituents. A few microphenocrysts of pyroxene are visible as well as masses of felted colorless amphibole and grains of magnetite, which probably are secondary after olivine. Pale-green actinolitic amphibole replaces some of the pyroxene, and a few scattered flakes of biotite are present.

Diabase with pegmatitic network.—In thin section the heterogeneous diabase in the lower part of the chill zone is composed of veinlets of coarser-grained (3.0–4.0 mm.) diabase in fine-grained diabase (Fig. 5b). Crystals of the coarse areas and finer-grained matrix interlock across the otherwise abrupt boundary. In no place does the coarse-grained diabase appear to be filling definite fractures in a finer rock.

In this zone the plagioclase is characteristically all more or less altered, especially in some of the coarser-grained areas. In addition, some interstitial micropegmatitic intergrowths of quartz and turbid alkalic feldspar are apparent in the coarser material. In the finer-grained diabase the composition of the plagioclase is An_{60-65}; that of the coarser network has normal zoning ranging from An_{55} to An_{30}. Two kinds of pyroxene, augite and pigeonite, are present. The augite is pale brown, and many crystals have a discontinuous development of (001) parting. Augite is partly replaced by uralitic amphibole, and some interstitial green chlorite is also present.

Opaque grains of ilmenite-magnetite are scattered through the rock and are especially prominent in the coarse material, where they form large skeletal or gridlike crystals interstitial to the plagioclase and pyroxene and in part intergrown with the pyroxene.

Normal diabase.—The normal diabase is very uniform in texture. Its grain size averages about 1 mm, increasing slightly in the lower part of the section. Pyroxene (augite and pigeonite) and plagioclase (average, An_{62}) are intergrown in typical diabasic texture, and some interstitial micropegmatite is visible (fig. 5a). The plagioclase is mostly fresh but all thin sections show at least some turbid alteration of parts of some crystals. Normal zoning of

[1] Identified in thin section by its translucence and marked pleochroism from dark brown to black. Polished surfaces are strongly pleochroic and anisotropic from blue to red in polarized light.

[2] True ophitic textures in which pyroxene is in excess and occurs in large plates enclosing laths of plagioclase are practically nonexistent in these rocks. Most have a subophitic (Krokström, 1933) or diabasic texture where the pyroxene fills the interstices between lath-shaped crystals of plagioclase (Johannsen, 1939, p. 207).

the plagioclase is common (An_{58-67}). Brownish augite with a characteristic fine parting parallel to (001) is intergrown in groups in which the individual crystal members of the mosaic have different optical orientations. Some colorless pigeonite in part inverted to grayish-green

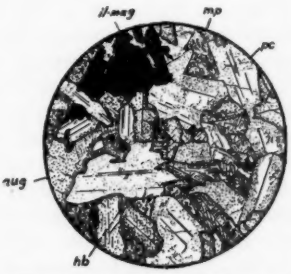

FIGURE 6. PEGMATITIC FACIES
184 feet below upper contact, drill hole 3 (pc, plagioclase; mp, micropegmatite; aug, augite; hb, hornblende; il-mag, ilmenite-magnetite) × 10.

hypersthene is also present. Skeletal ilmenite-magnetite crystals constitute an important minor constituent.

The downward transition from normal diabase to the pegmatitic facies is heralded by an increase in the amount of interstitial micropegmatite, severe alteration of the plagioclase, and development of rims of hornblende about the augite. The coarse pegmatitic facies comes in abruptly with little or no real gradation from the normal diabase.

Pegmatitic facies.—The pegmatitic facies differs from the normal diabase in its much coarser grain, hypidiomorphic texture, extensive alteration of the primary minerals, and larger amount of interstitial micropegmatite and coarse, skeletal crystals of ilmenite-magnetite (Fig. 6).

Purplish-brown augite in elongate crystals averaging 3 mm in length and as much as 7.5 mm long is the principal pyroxene; less than 5 per cent of pigeonite is present. The pyroxene crystals are mostly single individuals rather than intergrown groups. Many are twinned on (100); all have a well-developed (001) parting. The pyroxene has reaction rims of brown hornblende where it is in contact with the interstitial micropegmatite.

The original large plagioclase crystals (average length, 4 mm) are now mostly turbid with gray alteration products; only the central parts of the largest crystals are fresh and clear (An_{58}).

Ilmenite-magnetite is abundant in large skeletal crystals, some of which are intergrown

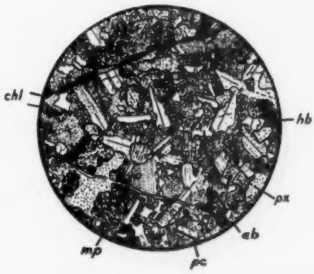

FIGURE 7. TRANSITIONAL GRANOPHYRIC DIABASE
From "Lower" Transition zone, 337 feet below upper contact; (pc, plagioclase; mp, micropegmatite; px, pyroxene; hb, hornblende; chl, chlorite; ab, albite) × 10.

with and molded on plagioclase and pyroxene. Much of the ilmenite-magnetite is accompanied by green to brownish-green stilpnomelane (?). Blebby grains of magnetite are also associated with the hornblende reaction rims around pyroxene.

Apatite is abundant as long, acicular prisms.

Transitional granophyric diabase.—(Fig. 7) The rocks from the transition zones above and below the granophyre more closely resemble diabase than granophyre. Texturally, they are similar to the normal diabase, and are composed of purplish-brown clinopyroxene intergrown with plagioclase, but contain no pigeonite. Interstitial micropegmatite is abundant and increases in volume toward the granophyre.

The plagioclase is mostly gray and turbid with alteration products, but the outlines of original large crystals are clearly visible. Some is clear and was determined as andesine (An_{45-47}). Reaction rims of brown and green hornblende border much of the pyroxene and locally wholly replace it. The pyroxene is partly bleached pale green and does not have the prominent (001) parting adjacent to the hornblende. Near the granophyre small amounts of darker-green pyroxene (hedenbergite) occur as elongate, ragged crystals in the micropegma-

tite. These appear to be the same as the pyroxene of the granophyre.

Ilmenite-magnetite as irregular, skeletal grains is plentiful. Acicular apatite is an abundant minor accessory. Substantial amounts of stilpnomelane (?)[3] and other chloritic materials

FIGURE 8. GRANOPHYRE

229 feet below upper contact, drill hole 3. (pc, plagioclase; q, quartz; hed, hedenbergite; ilv, ilvaite) × 10.

occur in irregular masses interstitial with the other minerals.

The change to true granophyre takes place within a distance of 4 feet. The amount of pyroxene rapidly diminishes, the brownish variety with the (001) parting gives way to the pale-green hedenbergitic pyroxene that is characteristic of the granophyre. One of the most important changes, of course, is the great increase of micropegmatite.

Granophyre.—As the name implies, this rock is granophyric, that is, the euhedral to subhedral feldspar and pyroxene are set in a matrix of micrographically intergrown quartz and feldspar (Fig. 8). Some of the quartz is not micrographically intergrown with feldspar but occurs as individual anhedral grains. The grain size is more variable than in the normal diabase, ranging from about 0.20 mm. in the finest-grained rock to nearly 0.40 mm. average grain size. The granophyre is finer-grained in general in the upper than in the lower part of the zone.

Feldspar, quartz, and hedenbergitic pyroxene are the principal mineral constituents. The euhedral plagioclase (An_{5-15}) is gray and turbid due to alteration. Some of it is optically continuous with the feldspar of the micropegmatite, which may completely surround a lath of plagioclase. Pyroxene is much less plentiful than in the diabase, amounting to about 5 per cent. Commonly the pyroxene has an elongate or plumose development, and some crystals have lengths a score or so times their width. Most are unaltered except for local replacement by hornblende or finely crystalline scaly green stilpnomelane (?) along their borders, and many of the crystals enclose some granules of magnetite.

Ilmenite-magnetite in irregular skeletal grains is the most prominent minor accessory. Associated with it is the hydrous calcium iron silicate, ilvaite, which is almost as abundant as the ilmenite-magnetite. Scaly greenish masses of stilpnomelane (?) are also common, occurring with the interstitial micropegmatite and the ilmenite-magnetite.

Besides these mineral accessories there are small amounts of acicular apatite and a few grains of epidote. Calcite, interstitial to all the other minerals, is present locally. A few grains of colorless zircon have been recognized. A nonmagnetic heavy mineral concentrate was prepared from the granophyre and examined microscopically. The concentrate contained a flood of apatite and some zircon crystals. The unbroken zircons all had sharp crystal terminations.

MINERALOGY

Olivine

Olivine is found only in the chilled contact zones of the diabase bodies. No concentration of olivine has been reported near the bottom of the Gettysburg sill like that near the base of the Palisades sill in New Jersey (Walker, 1940).

Alteration has completely destroyed the original olivine in the upper chill zone of the diabase beneath the sedimentary rocks at Dillsburg. Specimens from the chilled lower

[3] The following optical properties were determined in thin section and immersion liquids:
Color: dark green

$$(-)\ 2V = 0$$
$$Y \text{ and } Z = 1.655 \pm .002$$

Pleochroism: X, brownish yellow;

Z, dark green
$$X = 1.605 \pm .002$$
$$Z - X = .050 \pm .002$$

border of the main Gettysburg sill and the base of the offshoot from the main sill near Rossville, however, have relatively fresh olivine. It is reasonable to assume that the composition of these olivine crystals is the same as the altered olivines that crystallized early and were trapped in the upper chilled zone of the lower diabase at Dillsburg.

The "fresh" olivine microphenocrysts are irregular or rounded and have an average diameter of about 0.3 mm, ranging from 0.1 mm to 0.5 mm in diameter. They are colorless and unzoned. All are at least slightly altered to magnetite, chloritic material, and serpentine. Because of the scanty amounts and the small size of the crystals, no olivine was separated for refractive index determinations. The optic angle is negative and is constant at 87° ± 2°. According to Wager and Deer's diagram of variations in composition and optical properties of the olivine series (Wager and Deer, 1939b, p. 21), the composition is close to Fo_{80} Fa_{20}, the same as the olivine in the chilled zone at the bottom of the Palisades sill (Walker, 1940, p. 1068).

Feldspar

Plagioclase accounts for about 47 per cent of volume in the chilled border facies, from 45 to 50 per cent in the normal diabase, and about 18 per cent in the granophyre.[4]

The composition of the plagioclase becomes progressively more sodic in the drill-hole sequence. Average compositions range from An_{65} in the chilled facies, through An_{62} in the normal diabase, An_{55} in the pegmatitic facies, An_{47} in the transitional granophyric diabase, to An_{5-15} in the granophyre.

Micropegmatite

Micropegmatite is absent in the upper chilled zone; in the normal diabase, interstitial micrographic intergrowths constitute a small part of the rock. Micropegmatite is more abundant in rock from the transition zones and pegmatitic facies than in normal diabase, and in the granophyre it assumes the role of an essential constituent.

The feldspar of the micropegmatite is so fine-grained and clouded with submicroscopic inclusions that accurate optical data cannot be obtained. Patchy, discontinuous albite twinning and some pericline twinning can be seen under crossed nicols, and in many places the feldspar is optically continuous with earlier plagioclase crystals. It is probably a sodic plagioclase. Analyses (table 5) of the normal diabase and pegmatitic facies (which has an abundance of micropegmatite) show that Na_2O is clearly dominant over K_2O, suggesting that the feldspar is sodic plagioclase with a high content of K_2O.

TABLE 1.—QUARTZ-FELDSPAR RATIOS IN MICROPEGMATITE

Specimen	Per cent Quartz (volume)	Per cent Feldspar (volume)	Rock
333	45	54	Diabase with coarser network
409	45	55	Diabase
465	42	58	Pegmatitic facies
479	39	61	Transitional granophyric diabase
645	40	60	
560	39	61	Granophyre
601	40	60	
Average (volume)	41	58	

The ratio of quartz to feldspar in the micropegmatite intergrowths is very constant throughout the drill-hole section. In general (Table 1) the proportion is about 4 quartz to 6 feldspar.

Pyroxene

Particular attention was paid to pyroxene because of its importance as an accessory mineral throughout the diabase-granophyre sequence. Its composition, as indicated by variations in optical properties (Hess, 1949, p. 633–639), changes progressively toward an iron-rich variety in the succession diabase to granophyre. The optical data are summarized

[4] If the plagioclase intergrown with quartz in micropegmatite were included, the amount would be considerably higher. However, the micropegmatite intergrowths are considered as an individual mineral constituent itself.

in Table 2 and the compositions plotted in Figure 9.

Because of the scarcity of fresh pigeonite, no index data were obtained for it. Its optic angle is small and variable but averages about 18°.

Pigeonite, as such, is scanty. In thin section its former position is indicated by irregular greenish-gray, finely striated patches intergrown with the augite, a few with some clear unaltered pigeonite in them. The patches are

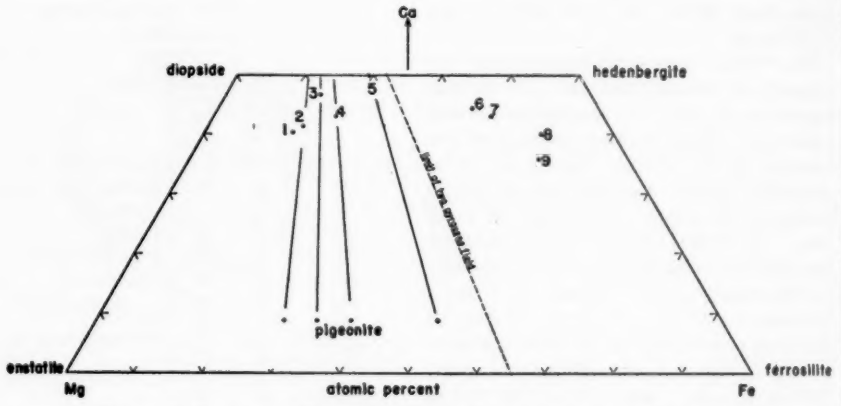

FIGURE 9. TRIANGULAR DIAGRAM SHOWING COMPOSITION OF THE PYROXENES

PYROXENE OF THE UPPER CHILL ZONE: The single variety of pyroxene in the upper-most microcrystalline chilled-contact facies of the diabase is colorless augite. Pigeonite, the clinopyroxene low in calcium, first appears in the lower part of the chill zone. Augite and pigeonite are intimately intergrown.

The augite is relatively rich in magnesium (see table 2, No. 2). The pyroxene from a specimen collected a short distance above the basal contact of the Gettysburg sill (No. 1) is somewhat lower in magnesium.

PYROXENE OF THE NORMAL DIABASE: The most abundant variety of pyroxene in the normal diabase is augite with a distinctly purplish-brown hue, suggesting that it contains a small amount of titanium. A finely-spaced parting parallel to (001) and twinning parallel to (100) are characteristic.

The augite (Nos. 3, 4) from this zone is somewhat richer in calcium and slightly richer in iron than pyroxene from the chill zone. The optic angles of these and other augites from the normal diabase are a little large and noticeably variable even in a single crystal, yet there is no visible zoning. The interior of the crystals have a smaller 2V than the outer part.

hypersthene containing microscopic lamallae of augite. The augite lamellae probably exsolved from pigeonite during cooling, and the pigeonite subsequently inverted to hypersthene (Hess, 1941, p. 580–581).

The augite is partly replaced by hornblende. The replacement is invariably accompanied by bleaching of the augite giving irregular, patchy areas of colorless to very pale-green pyroxene flecked with small plates of pale-green hornblende and devoid of the close-spaced (001) parting. The colorless augite is optically continuous with the purplish pyroxene, and apparently has the same index of refraction and optic angle.

PYROXENE OF THE PEGMATITIC FACIES: Most of the pyroxene in the pegmatitic facies of the diabase is a distinctly purplish-brown augite. There appears to have been only about 5 per cent or less original pigeonite.

Large euhedral crystals of augite are intergrown with plagioclase. Many of the augite crystals are elongated parallel to c, with length several times their width; the crystals are commonly twinned on (100) which, together with well-developed (001) parting, gives them a distinctive herringbone structure. The pi-

geonite that crystallized with the augite has inverted to hypersthene with exsolved lamellae of clinopyroxene, and appears as small greenish-gray areas within the brown augite, just as the pigeonite of the normal diabase.

apparent genetic relationship to the brownish pyroxene. The hedenbergite appears to come in abruptly and is not intergrown with the brown pyroxene, but appears to have crystallized somewhat later.

TABLE 2.—OPTICAL PROPERTIES OF CLINO-PYROXENES FROM THE DIABASE-GRANOPHYRE SEQUENCE

No. Specimen	Upper Chill Zone		Normal Diabase		Pegmatitic Facies	Transition Zones		Granophyre	
	1	2	3	4	5	6	7	8	9
	D-9-46	D3-296	D3-400	D3-440	D3-465	D3-645	D3-479	D3-601	D3-560
X	1.681	1.683	1.684	1.687	1.692	1.702	1.711	1.721	1.722
Y	1.685	1.687	1.690	1.693	1.699	1.707	1.720	1.728	1.729
Z	1.705	1.709	1.711	1.715	1.720	1.729	1.740	1.750	1.751
Z − X	0.024	0.026	0.027	0.028	0.028	0.027	0.029	0.029	0.029
Dispersion	r > v	r > v	r > v	r > v	r > v	r > v	r > v	r > v	r > v
Z ∧ C	42°	43½°	nd	nd	47°	nd	47°	45½°	49°
2V (+)	48°	49°	56° (zoned, 45° to 60°)	52° (zoned, 45½° to 59°)	58°	57°	57°	54°	57.5°
Atomic Per cent									
Ca	39	41	47	43.5	50	44.5	43	36	40
Mg	53	47	39	39.5	30.5	18.5	16.5	13	10.5
Fe	8	12	14	17.5	19.5	37	40.5	51	49.5

Hornblende replaces augite and forms a reaction rim between the pyroxene and interstitial micropegmatite.

The optic angle of the pyroxene from the pegmatitic facies (No. 5) is large (2V-58), which places its composition on the diopside-hedenbergite join (salite). It is therefore richer in calcium and iron than the pyroxene of the normal diabase, but lower in iron than similar-appearing pyroxene from the transition zone.

PYROXENE OF THE TRANSITION ZONES: Throughout most of the transition zone only a purplish-brown clinopyroxene of the composition of ferrosalite is found. Much of the pyroxene is partly replaced by hornblende and is bleached locally from brown to nearly colorless.

In transitional granophyric diabase a few feet from the granophyre, there are, in addition to the purplish-brown pyroxene, a few small, scattered crystals of hedenbergite which look like the hedenbergitic pyroxene in the granophyre. This clear-green pyroxene has no

PYROXENE OF THE GRANOPHYRE: Even in hand specimens, the pyroxene of the granophyre is obviously different from the augite of the diabase and pegmatitic facies of the diabase. Optical data indicate that it is hedenbergite. In thin section, the pyroxene is a pale-green, faintly pleochroic to nearly colorless, and characteristically has an elongate to plumose development. Many of the crystals are elongate, highly irregular, and poikilitically enclose earlier minerals (see Fig. 8). Some consist of several relatively small individual grains in optical continuity.

Some replacement by hornblende and, to a minor extent, biotite is apparent.

The pyroxene from the granophyre (Nos. 8 and 9) is relatively rich in iron, low in magnesium, but high in calcium. Like the pyroxene from the transition zones, 2V is large which, with the index of refraction, indicates a composition near the end member hedenbergite.

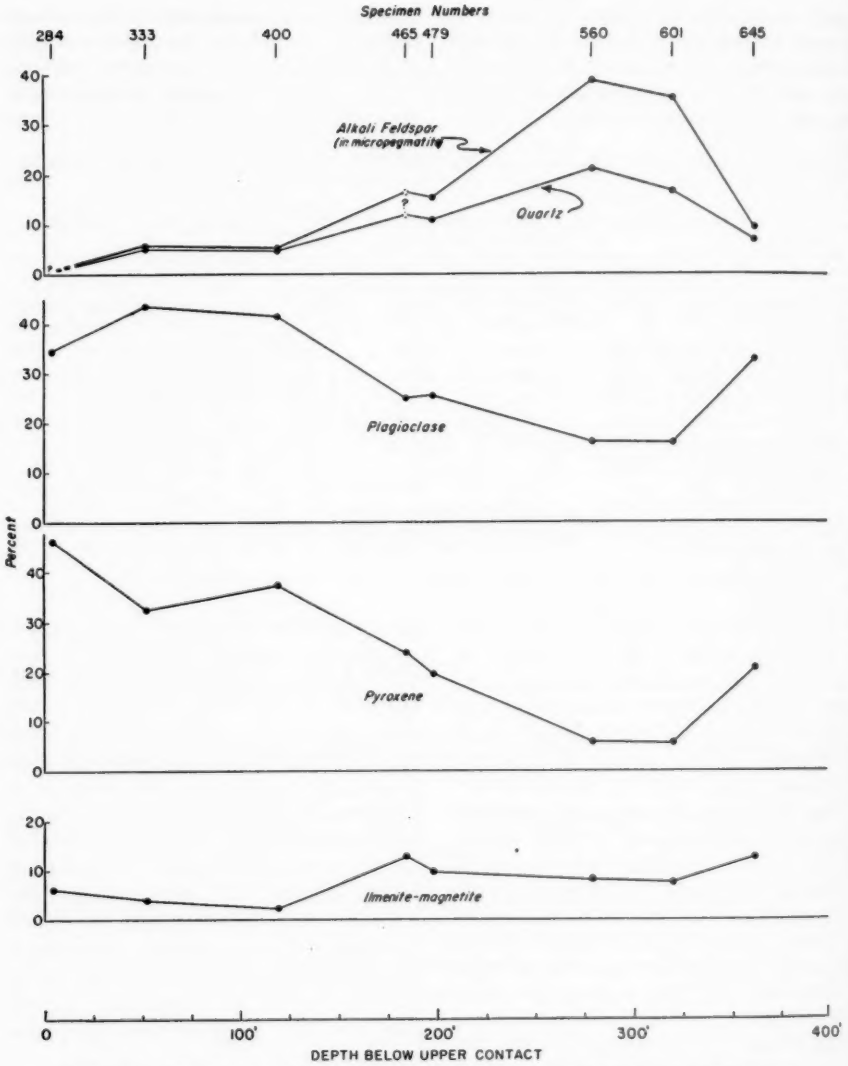

FIGURE 10. VARIATION IN MODAL MINERALS OF THE DIABASE-GRANOPHYRE SEQUENCE

Mineral Variation

The variation of the principal mineral constituents in the different zones of the diabase-granophyre sequence is shown in Figure 10, and the variation of modal minerals in analyzed specimens is presented in Table 3.

CHEMICAL DATA

Composition

To determine the variation in composition of the diabase-granophyre sequence, ten new rock analyses were made in the University of Minnesota laboratory by Mr. James Kerr and

TABLE 3.—MINERAL VARIATION IN THE DIABASE-GRANOPHYRE SEQUENCE

Zone	Plagioclase		Micropegmatite Per cent by Volume	Pyroxene	Ilmenite-magnetite
	Variation	Average			
Chilled zone	Most calcic observed = An_{70}	An_{65}	0–5	Augite only in uppermost part. Augite and pigeonite occur together beginning ca. 20 feet below contact	Small amounts of magnetite and ilmenite as small grains
Diabase with coarser network	Fine-grained matrix An_{60}–An_{55}	An_{62}	0–10	Augite and pigeonite. (Pigeonite largely inverted to hypersthene with clinopyroxene lamellae)	Magnetite with ilmenite lamellae; pure ilmenite
	Coarse-grained material An_{70}–An_{55}		12–20		
Normal diabase	An_{65}–An_{67}	An_{62}	12–15	Purplish-brown augite and pigeonite (mostly inverted)	Pure ilmenite as irregular intergrowths with magnetite containing ilmenite lamellae. Roughly equal amounts of pure ilmenite and ilmenite in magnetite
Pegmatitic facies		An_{55}	20–30	Purplish-brown clinopyroxene. (ferrosalite?) Inverted pigeonite	Abundant pure ilmenite and magnetite with ilmenite lamellae. Pure ilmenite in excess of magnetite with ilmenite lamellae. (ca. 60:40)
Transition zones	An_{55}–An_{47}	An_{45}	30–35	Purplish-brown clinopyroxene (ferrosalite?) No pigeonite or inverted pigeonite	Similar to occurrence in pegmatitic facies. In some of lower transition zone, ilmenite occurs almost exclusively. Ilvaite present
Granophyre	An_5–An_{15}		Av. = 30	Pale-green hedenbergite	Pure ilmenite and magnetite with ilmenite lamellae less abundant. Pure ilmenite predominant. Ilvaite common

Miss Eileen H. Kane (Table 4). Eight of the analyzed specimens are from core recovered from the drill hole that penetrated the "lower" or main diabase. The other two, 14 and 21, are, respectively, from the chilled bottom contact of the "upper" diabase at Dillsburg and the chilled contact facies at the bottom of the Gettysburg sill. The specimens from the drill core were chosen as representative of the principal varieties of diabase and granophyre. The modes, computed in weight per cent from Rosiwal analyses of thin sections of the analyzed specimens, generally show good agreement with the norm.

In Table 5, the chemical composition of the diabase at Dillsburg is compared with that of

TABLE 4.—CHEMICAL ANALYSES

Specimen Number	Diabase					Pegmatitic Facies	Transitional Granophyric Diabase		Granophyre	
	Chilled Facies			With Coarse Network	Normal					
	14	21	284	333	400	465	479	645	601	560
SiO_2	51.26	52.05	51.33	52.59	52.89	52.56	54.14	53.28	61.69	66.04
Al_2O_3	15.28	14.51	14.79	15.17	15.79	12.77	11.86	12.31	12.61	12.72
Fe_2O_3	1.66	1.37	1.85	2.22	1.71	4.78	3.08	3.43	2.98	2.48
FeO	8.98	9.00	8.32	7.85	9.19	10.18	12.57	12.27	8.32	6.55
MgO	7.31	7.30	7.23	5.21	4.33	2.69	1.92	2.61	0.77	0.54
CaO	10.18	10.61	10.51	8.85	9.78	7.21	5.69	6.91	4.04	2.65
Na_2O	2.06	1.91	1.94	2.74	2.49	3.65	3.21	2.88	5.71	4.62
K_2O	0.71	0.62	0.57	1.43	0.64	1.03	1.74	1.40	0.57	2.26
H_2O+	0.92	0.79	1.49	1.59	1.02	1.22	1.52	0.93	0.83	0.84
H_2O-	0.21	0.10	0.46	0.49	0.28	0.30	0.23	0.25	0.15	0.19
TiO_2	0.83	1.19	1.16	1.35	1.44	2.91	3.08	2.83	1.46	1.03
P_2O_5	0.10	0.14	0.14	0.17	0.18	0.27	0.37	0.31	0.50	0.22
MnO	0.21	0.18	0.17	0.15	0.18	0.20	0.24	0.22	0.15	0.11
S	0.08	0.03	nd	nd	nd	nd	nd	nd	nd	nd
Less O for S	−0.04	−0.01	—	—	—	—	—	—	—	—
Total	99.75	99.79	99.96	99.81	99.92	99.77	99.65	99.63	99.78	100.25
Sp. Gr.	2.97	2.98	2.94	2.93	2.94	2.99	2.89	2.95	2.80	2.72
FeO:MgO	1.23	1.23	1.15	1.50	2.12	3.78	6.54	4.70	10.80	12.13
Total Fe as FeO	10.47	10.23	9.98	9.85	10.73	14.48	15.34	15.36	11.00	8.78
Norms										
Quartz	2.08	4.02	4.02	4.20	6.84	8.00	10.35	9.65	15.23	21.69
Orthoclase	3.89	3.34	3.34	8.34	3.34	6.05	10.30	8.29	3.34	13.34
Albite	17.29	16.24	16.24	23.06	20.96	30.82	27.25	24.35	48.21	39.00
Anorthite	30.58	29.19	30.02	25.02	30.30	15.39	12.65	16.60	7.23	7.36
Diopside										
$CaSiO_3$	8.00	9.40	8.93	7.54	7.31	8.85	5.51	6.63	3.95	1.67
$MgSiO_3$	4.30	5.20	5.10	4.00	3.24	3.13	1.36	2.04	0.65	0.31
$FeSiO_3$	3.43	3.83	3.43	3.30	4.04	4.80	4.47	4.85	3.64	2.31
Hypersthene										
$MgSiO_3$	14.00	13.10	13.00	9.00	7.56	4.16	3.44	4.49	1.27	1.04
$FeSiO_3$	10.82	9.90	8.71	7.26	9.42	5.48	11.38	10.63	7.04	6.48
Magnetite	2.32	1.97	2.55	3.25	2.55	6.91	4.41	4.87	4.31	3.60
Ilmenite	1.52	2.28	2.13	2.58	2.74	5.53	5.93	5.38	2.74	1.98
Apatite	0.34	0.34	0.34	0.34	0.34	0.64	1.01	0.74	1.34	0.34
Water	1.13	0.89	1.95	2.08	1.30	1.52	1.75	1.18	0.98	1.03
Total	99.70	99.70	99.76	99.97	99.94	101.28	99.81	99.70	99.93	100.15
Normative plagioclase	An_{64}	An_{64}	An_{65}	An_{52}	An_{59}	An_{32}	An_{32}	An_{40}	An_{13}	An_{16}
Ab:An	0.565	0.557	0.541	0.921	0.691	2.00	2.15	1.47	6.67	5.30
$CaSiO_3$	19.7	22.7	22.8	24.2	23.5	33.5	21.0	23.2	23.9	14.1
$MgSiO_3$	45.1	44.2	46.2	41.8	34.2	27.6	18.4	22.8	11.6	11.4
$FeSiO_3$	35.2	33.1	31.0	34.0	42.6	38.9	60.6	54.0	64.5	74.5

TABLE 4 (*cont'd.*)

Specimen Number	Diabase					Pegmatitic Facies	Transitional Granophyric Diabase		Granophyre	
	Chilled Facies			With Coarse Network	Normal					
	14	21	284	333	400	465	479	645	601	560
Modes (in weight per cent)*										
Quartz		—	0.9	5.1	4.8	28.1	10.9	6.9	16.5	21.0
Alkali feldspar (in micropegmatite)		—		5.9	5.2		15.2	9.3	35.2	38.8
Plagioclase		30.0	34.4	43.7	41.5	25.0	25.6	32.8	16.0	16.2
Pyroxene		57.5	46.1	32.6	37.4	23.9	19.7	20.8	5.8	6.0
Olivine		2.0	—	—	—	—	—	—	—	—
Ilmenite-magnetite		7.0	6.2	4.1	2.5	2.9	9.9	12.6	7.6	8.2
Minor accessory mafics		3.2	12.4	8.3	8.4	9.8	18.1	16.9	18.2	9.5
Apatite		—	—	0.1	0.2	0.1	0.6	0.7	0.7	0.3
Sphene		—	tr.	—	—	0 2	—	—	—	tr.

* Determined by Rosiwal measurements using Wentworth integrating stage.

14 Chilled diabase at lower contact of upper diabase sheet. Core from drill hole 14, Dillsburg, Pa., Eileen H. Kane, analyst.

21 Chilled diabase within 1 foot of lower contact of Gettysburg sill, Beaver Creek, 0.4 miles from Conewago Creek, York County, Pa., Eileen H. Kane, analyst.

The following specimens are from core recovered from diamond drill hole 3, Dillsburg, Pa., analyses by James Kerr.

284 Chilled diabase 3 feet from upper contact of lower diabase.

333 Fine-grained diabase with coarser-grained network 52 feet below upper contact.

400 Normal diabase 119 feet below upper contact.

465 Pegmatitic facies of diabase 184 feet below upper contact.

479 Transitional granophyric diabase above granophyre 198 feet below upper contact.

645 Transitional granophyric diabase beneath granophyre 364 feet below upper contact.

601 Granophyre near lower transition zone 320 feet below upper contact.

560 Granophyre from central part of granophyre zone 279 feet below upper contact.

the Gettysburg sill and other diabase bodies of Triassic age from the eastern United States. Analyses of diabase intrusives of similar composition from outside the United States are also included for comparison. All analyses are of specimens from the fine-grained, chilled contact of intrusions and hence represent as closely as possible the composition of the original undifferentiated magma. All the analyses listed are of the saturated or tholeiitic magma type of Kennedy (1933).

The close general similarity in composition of these representatives of a magma type having world-wide distribution is remarkable. There is great uniformity of composition between the diabase of the Dillsburg and Gettysburg areas and the diabase that forms the New York Palisades as might be expected, for both groups are members of the same province and are identical in age and mode of occurrence.

In Table 6, the chemical composition of the granophyre from the drill hole is compared with granophyres of similar composition from other localities. There is some variation in composition but all analyses have one feature in common, the Na_2O content is in excess of the K_2O content.

TABLE 5.—COMPARISON OF DILLSBURG DIABASE WITH ROCKS OF SIMILAR MAGMA TYPE

	Dillsburg		Gettysburg sill	Palisades sill		Whin sill	Average Karoo	Downes Mountain	Average Tasmanian dolerite
	1	2	3	4	5	6	7	8	9
SiO_2	51.33	51.26	52.05	52.12	51.82	50.72	52.5	53.21	52.65
Al_2O_3	14.79	15.28	14.51	14.33	14.76	13.76	15.4	13.95	16.23
Fe_2O_3	1.85	1.66	1.37	1.34	1.22	3.87	1.2	1.24	0.51
FeO	8.32	8.98	9.00	8.86	9.24	8.50	9.3	8.93	8.21
MgO	7.23	7.31	7.30	7.76	7.32	5.42	7.1	7.19	6.64
CaO	10.51	10.18	10.61	10.55	10.02	9.09	10.3	8.95	11.34
Na_2O	1.94	2.06	1.91	2.01	2.06	2.42	2.1	2.65	1.58
K_2O	0.57	0.71	0.62	0.78	0.82	0.96	0.8	1.13	0.90
H_2O+	1.49	0.92	0.79	0.75	0.84	1.51	—	0.49	0.48
H_2O-	0.46	0.21	0.10	0.16	0.14	0.76	—	0.22	0.85
TiO_2	1.16	0.83	1.19	1.14	1.34	2.39	1.0	1.73	0.58
P_2O_5	0.14	0.10	0.14	0.14	0.12	0.26	0.1	0.21	0.01
MnO	0.17	0.21	0.18	0.18	0.12	0.16	0.2	0.23	0.15
S	—	0.08	0.03	—	—	—	—	—	—
	99.96	99.79	99.80	100.12	99.82	99.82	100.00	100.13	100.13
Less O for S	—	0.04	0.01	—	—	—	—	—	—
		99.75	99.79						

1. Three feet below upper contact of "lower" diabase; core from drill hole 3, Dillsburg, Pa., specimen 284, Table 4, James Kerr, analyst.
2. Chilled facies at bottom of "upper" diabase sheet; core from drill hole 14, Dillsburg, Pa., specimen 14, Table 4, Eileen H. Kane, analyst.
3. Chilled facies within one foot of lower contact of Gettysburg sill, Beaver Creek, 0.4 miles from Conewago Creek, York County, Pa., specimen 21, Table 4, Eileen H. Kane, analyst.
4. Average of 6 analyses of chilled border facies of diabase, Palisades sill, New Jersey. (H. H. Hess, unpublished data)
5. One foot above lower contact, Palisades sill, New Jersey, F. A. Gonyer, analyst (Walker, 1940, p. 1080).
6. Whin Sill, England (average of 6) Harwood, analyst. (Holmes and Harwood, 1928, p. 539).
7. Average Karoo dolerite, South Africa. (Walker and Poldervaart, 1949, p. 649)
8. Downes Mountain dolerite, Calvinia, South Africa. A. J. Hall, analyst. (Walker and Poldervaart, 1941, p. 171.)
9. Average undifferentiated Tasmanian dolerite (six analyses). A. B. Edwards, analyst. (Edwards, 1942, p. 465.)

Chemical Variation

Figure 11 illustrates the variation in chemical composition of the diabase-granophyre sequence at Dillsburg. This method of plotting the components using depth below the upper contact as the abscissa was found to be more satisfactory than the usual method where silica is used as the base. The corresponding changes in weight per cent of the modal minerals are shown in Figure 10.

A slight difference in chemical composition between the chilled contact facies (284) and the underlying normal diabase (400) is apparent. The normal diabase is a bit more silicic because a somewhat longer cooling period of the diabase magma beneath the chill zone permitted slight fractionation. Alumina reaches a maximum in the normal diabase (400), suggesting a slight relative enrichment in plagioclase resulting from some sinking and removal of pyroxene in the beginning stages of crystal-

lization. This assumption is consistent with the mode, which has a relatively high plagioclase content, though (333) has slightly more plagioclase than (400).

A slight break in the curves between the chilled zone (284) and normal diabase (400) is given by the diabase with the coarser-grained network in fine-grained matrix (333). The coarser pegmatitic material of this rock (333) with its sericitized somewhat more sodic plagioclase and abundant interstitial micropegmatite gives a lower total CaO and Al_2O_3 content, but raises the amount of alkalies and silica.

A major break in the curves comes in the pegmatitic facies of the diabase (465) and in the transition zones (479, 645). Silica increases abruptly and rises considerably in relative amount in the granophyre (560, 601), as would be expected from the increase of quartz in the mode. In the transition zones (479, 645) and granophyre (560, 601), Al_2O_3 falls considerably below its amount in the normal diabase (400), and is reflected in the mode by a corresponding change in the content and composition of the plagioclase.

MgO decreases steadily from the chilled diabase toward the granophyre, which has but a small amount of low-magnesia pyroxene. The decrease of CaO in the same direction reflects the increasingly sodic composition of the feldspar.

Analyses of the rocks from the intermediate zones (479, 645) reveal a high content of total Fe accompanied by a similar enrichment in TiO_2. The norms show a concomitant increase of magnetite and ilmenite, which is corroborated by an abundance of ilmenite-magnetite in the modes. In the granophyre (560, 601) total iron is relatively very low. There is an absolute enrichment in iron in the transition zones (479, 645) and an increase in the FeO:MgO ratio; but in the granophyre (560, 601), which shows a great increase in FeO:MgO, there is actually a marked decrease in absolute iron. If the granophyre is a differentiate from the diabase magma, an important point in petrogenesis is demonstrated, namely, iron is concentrated in the magma at a late stage in its crystallization history, but not in the final residual product (Fig. 12).

Figure 11 and more clearly Figure 13 illustrate the change from high calcium and low potassium in the chilled facies and normal diabase to rock containing more alkalies and less CaO in the granophyre. It is interesting to note (Table 4) that although the proportion of K_2O increases in the granophyre, it does not exceed the Na_2O content. This is in accordance with Bowen's conclusion (1928, p. 98, 104) that, with limited reaction (strong fractionation) in the plagioclase feldspar series, the Na_2O content increases in the residual liquid and remains high in the final stage so that, although K_2O likewise increases, it does not attain or exceed the amount of Na_2O.

The Na_2O:K_2O ratios of 21 analyzed granophyres including the granophyre near Dillsburg are listed in table 7 and plotted in Figure 14. A general correlation between the size of the granophyre body and Na_2O:K_2O ratio is

TABLE 6.—COMPOSITION OF THE DILLSBURG GRANOPHYRE COMPARED WITH OTHER GRANOPHYRES

	Dillsburg 560	A	B	C	D
SiO_2	66.04	65.20	64.13	58.81	63.33
Al_2O_3	12.72	13.72	13.15	12.02	12.99
Fe_2O_3	2.48	3.63	1.08	5.77	2.87
FeO	6.55	3.72	6.31	9.38	5.12
MgO	0.54	1.01	1.08	0.72	2.21
CaO	2.65	2.79	3.62	5.03	2.05
Na_2O	4.62	5.22	3.64	3.91	3.97
K_2O	2.26	2.17	2.32	2.39	3.52
H_2O^+	0.84	1.27	2.71	0.21	1.25
H_2O^-	0.19	0.72	0.36	0.19	0.58
TiO_2	1.03	0.39	1.19	1.26	1.55
P_2O_5	0.22	0.38	0.31	0.71	0.37
MnO	0.11	—	0.27	0.21	0.16
BaO	—	—	0.09	—	—
Total	100.25	100.22	100.26	100.61	99.97

A. "Red veinlet" in Whin Sill. Tomkeieff (1929), Analysis III

B. Inninmorite pitchstone. Bailey et al. (1924, p. 19)

C. Hedenbergite granophyre 3047, Skaergaard intrusive. Wager and Deer (1939a), p. 210.

D. Red rock, Northland Sill, Duluth. Schwartz and Sandberg (1940), p. 1144.

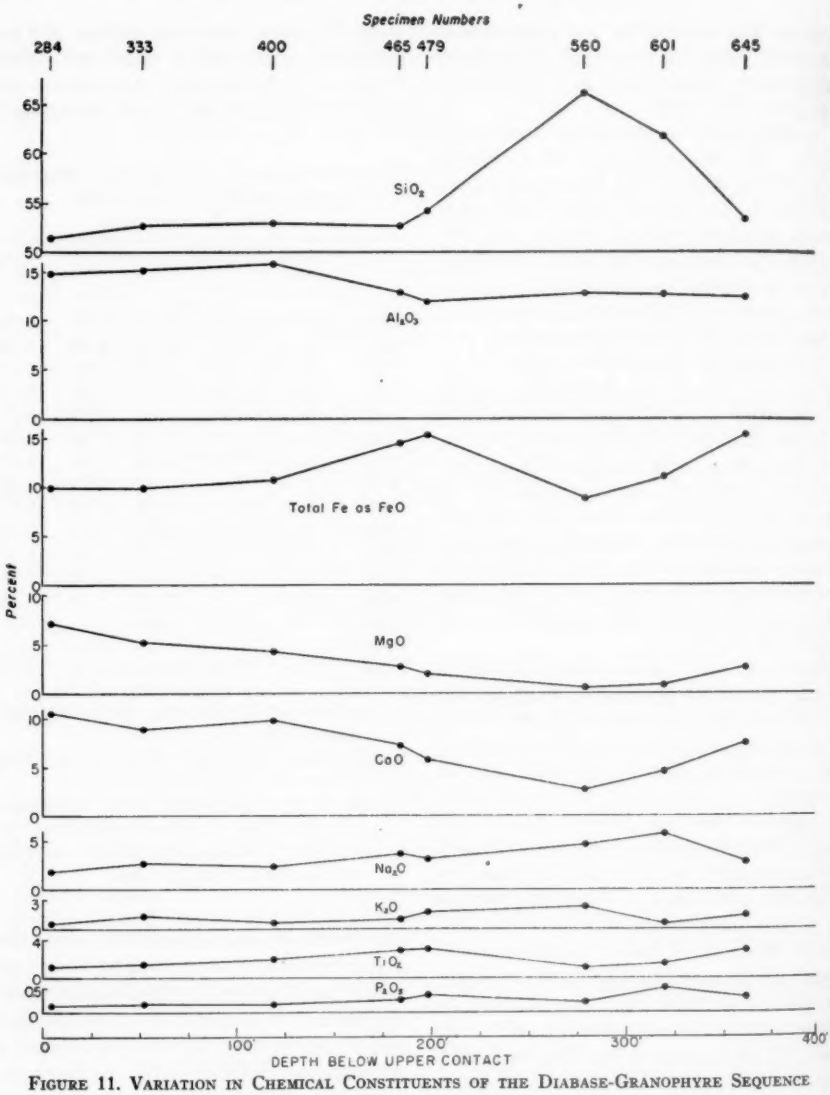

FIGURE 11. VARIATION IN CHEMICAL CONSTITUENTS OF THE DIABASE-GRANOPHYRE SEQUENCE

apparent from the table.[5] Of the 10 analyses having Na_2O in excess of K_2O, 6 are from definitely small bodies, while 9 of the 11 bodies with K_2O greater than Na_2O are known to be large. Exceptions to this general relationship can be found, as in the case of the analyses of the Carrock Fell granophyre (F) and the Knock Ring dike (J) in which Na_2O exceeds K_2O, or the small micropegmatite dike in the Karroo (S) where K_2O is greater than Na_2O.

[5] It is difficult to obtain accurate data on thickness and extent of most of the granophyre bodies for which analyses are given. By "small" is meant bodies ranging from veins and dikes a few inches wide to bodies 100 feet or so thick. "Large" bodies include those from around 100 feet to several hundred feet thick and usually with a considerable horizontal extent.

TRACE ELEMENTS: Mr. K. J. Murata of the U. S. Geological Survey made spectrographic determinations of trace elements in the analyzed rocks (Table 8). For the most part, the distribution of the trace elements is in accordance with expectations. The discussion that follows relies heavily on the important results of Wager and Mitchell's work (1951, esp. pp. 177–196) on the distribution of trace elements in rocks of the Skaergaard intrusion, East Greenland. The distribution of trace elements in rocks of the Dillsburg diabase-granophyre sequence follows closely the pattern displayed by the same elements in rocks of the Skaergaard intrusion. The elements are discussed, as they are arranged in Table 8, in order of size of the ionic radii.

The presence of boron in one of the specimens of granophyre is puzzling, particularly as it has not been found in the other analyzed granophyre. No boron mineral has been recognized in this specimen, but the rock contains considerable hornblende and chloritic material that may contain the element. Rankama and Sahama (1950, p. 486) state that . . . "boron, like beryllium, is an element characteristic of the late stages of magmatic crystallization," . . . and that (p. 487) "the bulk of boron is enriched in residual magmas and solutions. . . ."

Beryllium near the limit of sensitivity is found in the transitional granophyric diabase and in the granophyre, but was not detected in the diabase and pegmatitic facies. The occurrence is probably due to the tendency for beryllium to become enriched toward the late stages of magmatic differentiation. Probably the element is contained in the alkalic feldspar of the late crystallizing facies (Rankama and Sahama, 1950, p. 443, 444).

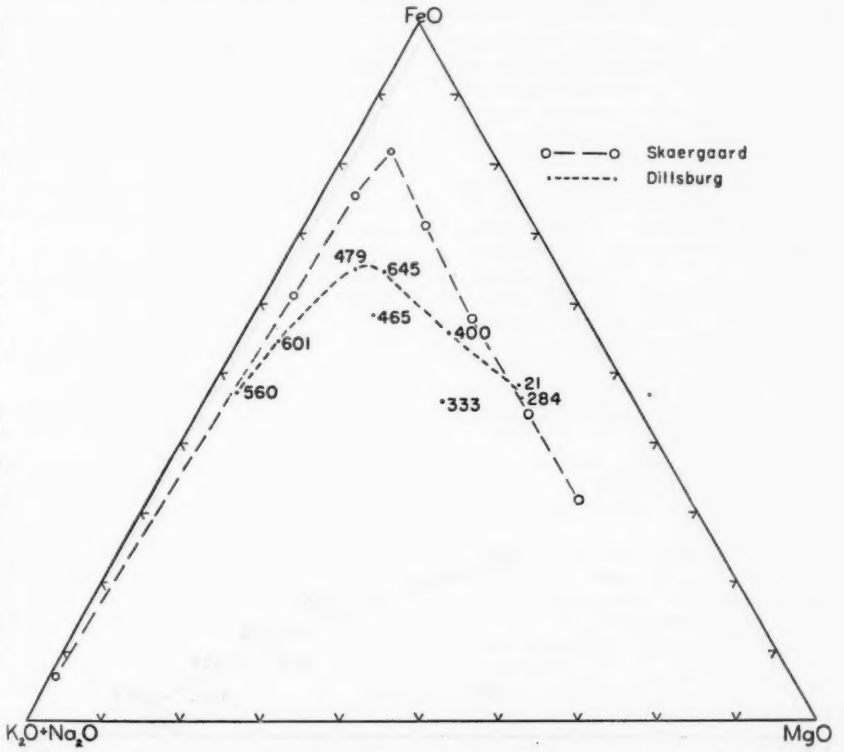

FIGURE 12. VARIATION OF FeO, MgO, AND $K_2O + Na_2O$ IN THE DIABASE-GRANOPHYRE SEQUENCE

Chromium is most abundant in the chilled facies, which approximates the composition of the original magma. The concentration falls off rapidly. In the normal diabase it is but one-tenth, and, in the pegmatitic facies, one one-hundredth of the value in the chilled facies. Chromium was not detected in the granophyre and rock intermediate between granophyre and diabase. Probably, as found by Wager and Mitchell (1951, p. 150, 183–184), the chromium is incorporated in the early clinopyroxene, and so the remaining magma is progressively depleted of chromium by fractional crystallization until in later stages it is effectively absent. Unlike the latest granophyric differentiates of the Skaergaard in which chromium reappears in small amount, no chromium is found in the granophyre at Dillsburg.

Vanadium is constant in the chilled zone and normal diabase, and attains its highest concentration in the pegmatitic facies. None was recorded in the specimens of granophyre. Vanadium is a constituent of pyroxene and, as shown by Wager and Mitchell's work (1951, p. 185–186), has a distinct preference for magnetite. Accordingly, the greatest concentration in the diabase near Dillsburg is in the pegmatitic facies, which contains abundant magnetite; specimen (645) from the transition zone likewise contains more magnetite than the diabase and the other specimen of granophyric diabase (479). The absence of vanadium from the granophyre is due to the vanadium having been essentially removed from the magma by the time the final residuum crystallized.

Nickel is most abundant in representatives of the chill facies of the diabase. The element decreases in the progressively more felsic repre-

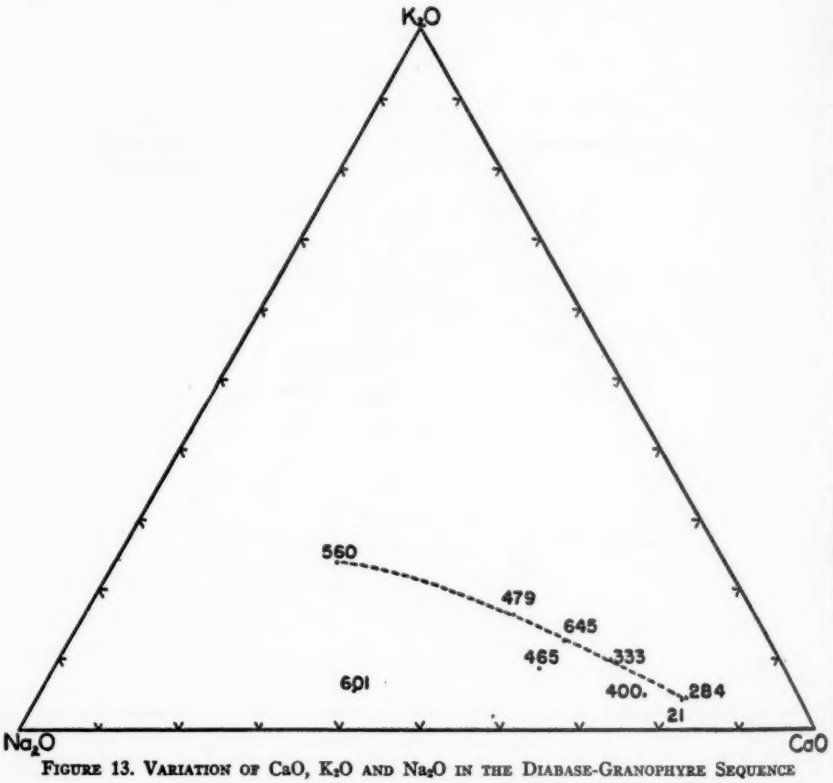

FIGURE 13. VARIATION OF CaO, K_2O AND Na_2O IN THE DIABASE-GRANOPHYRE SEQUENCE

TABLE 7.—ALKALIES IN GRANOPHYRES, AND COMPARISONS BETWEEN SIZE OF BODIES AND SODA-POTASH RATIO

		Per cent Na_2O	Per cent K_2O	Ratio Na_2O: K_2O	Size of body	Location
Na_2O greater than K_2O	A	4.62	2.26	2.04	Small	Dillsburg, No. 560 (this paper)
	B	5.22	2.17	2.40	Small	Red veinlet in Whin Sill. Tomkeieff (1929)
	C	3.91	2.39	1.64	Small	Hedenbergite granophyre 3047, Skaergaard intrusive. Wager and Deer (1939a), p. 210
	D	4.24	3.85	1.10	Small	Acid granophyre 3058, Skaergaard intrusive Wager and Deer (1939a), p. 208
	E	4.27	3.06	1.40	Small	Quartz-porphyry, Rum. Harker, (1908) quoted in Wager and Deer (1939a), p. 208
	F	5.55	3.53	1.57	Large	Augite-granophyre, Carrock Fell, Cumberland. Harker (1895), quoted by Wager and Deer (1939a), p. 208
	G	3.97	3.52	1.13	Small (?)	Red-rock, Northland sill, Duluth. Schwartz and Sandberg (1940), p. 1144
	H	2.48	2.37	1.05	Small	Acid rock, Moyie Sills, British Columbia. Daly (1905)
	I	5.12	4.08	1.25	?	Augite-granophyre, major intrusion, center 2, Ardnamurchan. Richey and Thomas (1930), p. 84
	J	3.92	2.34	1.67	Large (?)	Augite-granophyre, Knock Ring dike, Mull. Bailey et al., (1924) quoted by Wager and Deer (1939a), p. 208
K_2O greater than Na_2O	K	3.90	4.67	0.83	Large	Granophyre, Beinn a' Ghraig ring dike, Mull. Bailey *et al.* (1924) p. 20
	L	3.61	4.90	0.74	Large	Hornblende-granophyre, Skye. Harker (1904), quoted by Wager and Deer (1939a), p. 208
	M	4.24	4.52	0.94	Large (Separate intrusion?)	Granophyre, Breven dolerite dike. Krokström (1932), p. 35
	N	3.40	4.10	0.83	Large	Red-rock, Endion Sill, Duluth. Schwartz and Sandberg (1940), p. 1144
	O	2.82	4.07	0.69	Large	Red-rock, Lester River Sill, Duluth. Schwartz and Sandberg (1940), p. 1144
	P	3.45	3.98	0.87	Large	Red-rock, Duluth gabbro. Grout (1918) p. 65
	Q	3.39	3.82	0.89	Large	Average Sudbury micropegmatite. Collins (1934)
	R	2.40	5.20	0.46	Large	Granophyric granite, Bushveld. Hall (1932), p. 375
	S	4.29	5.58	0.77	Small	Microgranitic dike, Insizwa. Scholtz (1936), p. 143
	T	3.44	4.97	0.69	Large (?)	Red rock, Pigeon Point. Minn. Grout (1918) p. 653
	U	4.15	4.47	0.93	Large (?)	Granophyre, Glen More ring dike, Mull. Bailey *et al.* (1924), p. 29

sentatives of the diabase-granophyre sequence, and is absent from the granophyre. It is also absent from one of the two specimens of transitional granophyric diabase (479) and present in minimum concentration in the other (645).

Nickel substitutes for magnesium and ferrous iron in olivine, pyroxene, ilmenite, and magnetite. Olivine is found only in the chilled facies of the diabase, so the nickel must be contained in the pyroxene, ilmenite, and magnetite in the

other facies. There is a fair correlation between the nickel content and amount of modal pyroxene. As Wager and Mitchell (1950, p. 187–188) emphasize, the reduction of the nickel content in the later differentiates is not due to difficulty in entry of the element in minerals of later fractions but to depletion of nickel in the liquid by removal in earlier fractions.

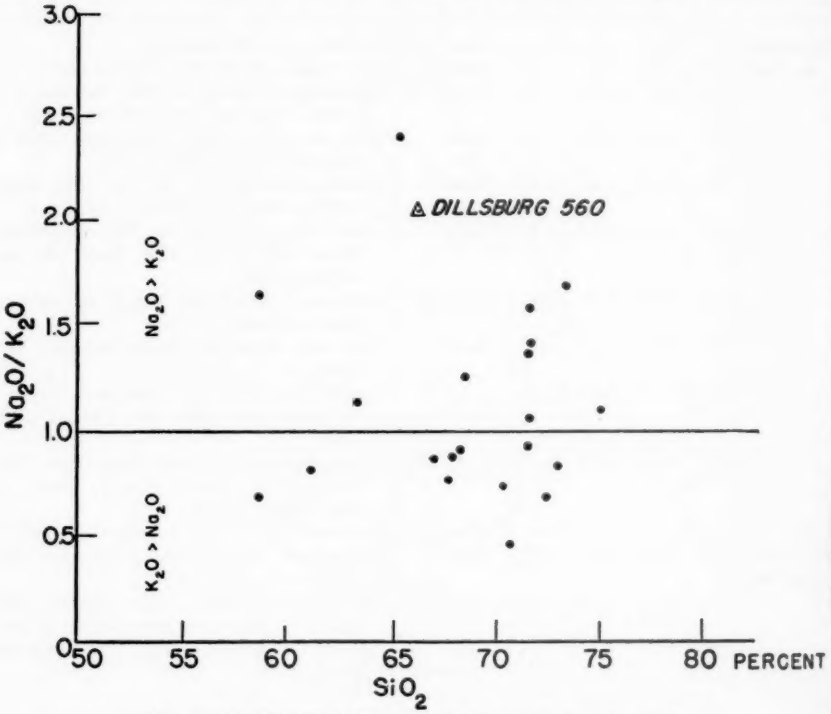

FIGURE 14. Na_2O:K_2O RATIO OF 21 ANALYZED GRANOPHYRES

The amount of cobalt is very constant from the chilled facies of the diabase, through the transitional granophyric diabase, and is slightly lower in the specimens of granophyre. Cobalt is only about a third as abundant as nickel in the original magma, represented by the chilled facies, but it does not decline as rapidly as nickel in the more felsic facies, though cobalt enters the same minerals as nickel (Wager and Mitchell, 1951, p. 150–158, 189).

The validity of the distribution of copper is uncertain because the samples used for spectrographic analysis were splits of the powders used for chemical analysis, and these powders were sifted through brass screens. The amount of copper rises in specimens of the transitional granophyric diabase and granophyre, and there appears to be some direct relation between high iron content and copper (479), though copper is highest in (601), which has about the same amount of iron as rock in the chilled zone and normal diabase. Possibly the copper content of the magma rose slightly during fractional crystallization because only a part of it was removed in the silicate minerals and magnetite, so that the final residue which crystallized to form the granophyre had a slightly higher concentration of copper than the original magma. Unfortunately, sulfur was not determined in the chemical analyses, so a comparison between copper and sulfur cannot be made.

Zirconium is least abundant in the chilled facies of the diabase and only slightly more abundant in the normal diabase. The element

is more abundant in the pegmatitic facies and the transitional rock, and attains its greatest concentration in the granophyre. The gradual increase of zirconium in the progressively more felsic rocks indicates an increasing content of zirconium in the liquid from which these rocks crystallized. Rankama and Sahama (1950, p. 565) say that zirconium is not readily incorporated in mineral structures, but Wager and Mitchell (1951, p. 151–159, 192) found that it entered the pyroxene and apatite, proxying for calcium. They found small amounts of zirconium in the early and middle pyroxenes, but it was absent in later ones. Apatite is fairly abundant in the granophyre, but only a few minute grains of zircon have been seen. Presumably, though, where zirconium attains higher concentrations, it tends to precipitate as zircon. Most of the zirconium in the later differentiates at Dillsburg is in the apatite.

Yttrium is present throughout the diabase granophyre sequence and is concentrated in the granophyre and granophyric diabase. Lanthanum is concentrated in detectable amounts in the rocks that contain the most yttrium, though it is below the sensitivity in the diabase and pegmatitic facies. Both elements enter apatite abundantly and pyroxene in moderate amounts (Wager and Mitchell, 1950, p. 193–194).

Strontium is consistently the most abundant trace element in the rocks at Dillsburg. Its variation is not clearly understood, but apparently it tends to be concentrated in the transitional rocks and pegmatitic facies; it is most abundant in the diabase with coarser, pegmatitic network (333), and is least abundant in the granophyre (560). Strontium replaces calcium in plagioclase and apatite (Wager and Mitchell, 1951, p. 194–195).

Barium is least abundant in the chilled facies, and more abundant in the normal diabase and pegmatitic facies. It increases slightly in the granophyric diabase and attains its maximum in the granophyre (560), though it is low in one of the granophyre specimens (601) which is somewhat less felsic and has an abnormally small amount of potash. The variation of barium is fairly well correlated with potash. Probably some of the barium is in the potassic feldspar of the interstitial micropegmatite though Water and Mitchell (1951, p. 149, 195) found that in the Skaergaard rocks barium entered plagioclase. The increase of barium in later rocks is probably due to an increase of barium in the magma, for Wager and Mitchell

TABLE 8.—TRACE ELEMENTS

Specimen	Diabase				Pegmatitic facies	Transitional granophyric diabase		Granophyre		Ionic radii in A°
	Chilled facies		With coarse network	Normal						
	21	284	333	400	465	479	645	601	560	
Be	—	—	—	—	—	0.000x	0.000x	0.000x	0.000x	0.34
B	—	—	—	—	—	—	—	0.007	—	0.20
Cr	0.02	0.02	0.008	0.002	0.0002	—	—	—	—	0.64
V	0.025	0.025	0.025	0.025	0.06	0.0075	0.03	—	—	0.65
Ni	0.009	0.0085	0.006	0.0045	0.0035	—	0.002	—	—	0.78
Co	0.0035	0.0035	0.0035	0.0035	0.004	0.0035	0.004	0.002	0.0015	0.82
Cu	0.02	0.015	0.015	0.015	0.02	0.035	0.025	0.04	0.025	0.83
Zr	0.006	0.0055	0.007	0.007	0.01	0.02	0.015	0.025	0.030	0.87
Y	0.0025	0.002	0.003	0.003	0.006	0.0095	0.007	0.0095	0.010	1.06
La	—	—	—	—	—	0.00x	—	0.00x	0.00x	1.22
Sr	0.04	0.04	0.06	0.04	0.05	0.05	0.045	0.04	0.025	1.27
Ba	0.015	0.01	0.02	0.02	0.02	0.025	0.025	0.01	0.035	1.43

Looked for but not found: Pt, Au, Pb, Sn, Bi, Sb, As, Ge, Tl, In, Zn, Mo. K. J. Murata, U. S. Geological Survey, analyst.

found that the ease of entry in the feldspars was fairly constant throughout the rock series.

Petrology

General Statement

The magma that rose into openings in the crust and solidified as diabase or flowed out on the surface as basalt in southeastern Pennsylvania and elsewhere in the areas of Triassic rocks in the eastern United States was unquestionably a true primary magma. A primary magma is indicated by the widespread distribution and enormous volume represented, the uniform composition of its chilled phase, and the occurrence of basalt and diabase without complementary igneous rocks. The magma must have risen practically simultaneously throughout the region and moved rapidly into place, for the uniformity of composition of its chill facies everywhere means that the magma was emplaced while still in an almost wholly liquid state before it had time to differentiate. As it solidified, certain processes of differentiation within this primary magma formed a small volume of late-stage rock types differing markedly in composition from the original magma.

Origin of the Diabase-Granophyre Series

General statement.—According to Bowen (1928, p. 72) and Kennedy (1933), magma of the composition that formed the intrusions of Triassic age diabase, though showing silica in the norm, is not so siliceous that it cannot precipitate out olivine as an early phase. In composition, it lies very near the clinopyroxene-forsterite boundary but within the forsterite field of the investigated system anorthite- forsterite-silica (Bowen, 1928, p. 42, 72). Thus, in the chilled contact facies of the diabase, we find some microphenocrysts of olivine that formed due to slight cooling of the magma as it was intruded and which were trapped by freezing of the liquid at the contact.

Crystallization of plagioclase and pyroxene commenced soon after olivine. The petrographic data are inconclusive as to whether pyroxene or plagioclase was the first to crystallize, but in the chilled zone at the bottom of the Gettysburg sill microphenocrysts of augite, some of them enclosing one or two small plagioclase laths, are frozen in a fine-grained groundmass composed of plagioclase and intergranular augite.

Because heat was lost more rapidly from the upper part of the body, a relatively fine-grained roof of diabase was formed. Complete solidification took place in the upper part before the plagioclase and pyroxene could change much in composition; so we find a relatively fine-grained uniform diabase with relatively magnesia-rich pyroxene and zoned plagioclase of a composition not much more sodic than that which formed at the outset of crystallization. In the deeper, still fluid parts of the intrusion, crystallization was proceeding more leisurely, and we may suppose by analogy with other similar intrusions that, there was a certain amount of gravitative settling of pyroxene and some early olivine.

By the time crystallization had progressed to the "intermediate" stage, there remained a residual liquid that in part occupied crystal interstices and in part was concentrated locally in the upper part of the body as pockets of liquid. The writer is not prepared to specify what mechanism brought about the local concentration of these pockets of residual liquid. Crystallization from the top and bottom of the intrusion, leaving a residual liquid between essentially solidified diabase, was possibly the dominant process of concentration, though other factors, including gravity and convection, may have contributed in part. The concentration of iron and titania, alkalies, silica, and water in the remaining liquid was high, while magnesia and lime had been depleted. Correspondingly, we find the plagioclase more sodic and the pyroxene relatively rich in iron, and an abundance of alkali feldspar and quartz as micropegmatite between the pyroxene and plagioclase crystals. A good deal of the iron and titania combined to form titaniferous magnetite and ilmenite. The lower temperatures and high water content were favorable for the formation of some hornblende, which formed by reaction between the pyroxene and the magma. A little biotite also crystallized. The formation of hornblende and biotite released still more silica to the remaining liquid.

Mineralization.—The intermediate stage was critical in so far as the Dillsburg magnetite district was concerned, for now the magma rich in iron was at the same time charged with volatiles. Here was a potential mineralizing agent that could enter the country rocks, metamorphose them, and deposit iron as magnetite.

The residual magma of the intermediate stage had collected below the solid, impervious cover of diabase. The cover was fractured, perhaps by movements in the underlying partly fluid, partly solid body brought about by surges of fresh magma in depth; perhaps by the high internal pressures being developed by crystallization; or perhaps merely by cooling of the cover and the formation of tension cracks. Some of these openings reached into the portion in which crystallization was fairly well advanced. Under pressure, the volatile-rich solutions escaped, taking with them some of the oxides, particularly iron. These hydrothermal solutions passed through the overlying sedimentary rocks, altering them and forming new minerals and, finding the layers of limestone conglomerate, replaced the calcium carbonate with new minerals, among which were magnetite and calcium-iron silicates.

Pegmatitic facies.—The pegmatitic facies of the diabase represents magmatic liquid entrapped in essentially solidified, but possibly still mushy, diabase at a stage when crystallization of much of the diabase had enriched the liquid in viscosity-reducing volatiles as well as alkalies and iron. From this liquid, plagioclase and pyroxene continued to crystallize with coarser grain than in the surrounding, essentially solidified diabase.

With continued crystallization, the remaining liquid changed in composition and toward the end was no longer in equilibrium with the solid phase. Thus the plagioclase is strongly zoned and shows continuous growth into the alkali feldspar intergrown micrographically with quartz; the final solution altered the earlier formed plagioclase. Likewise, the changing solutions reacted with the pyroxene and formed rims of hornblende and some biotite. The removal of iron as magnetite rather than in combination with silica as pyroxene, and the formation of hornblende and biotite, augmented the amount of silica in the remaining liquid and thereby increased the quartz in the final interstitial material.

The heterogeneous diabase with its coarser-grained patches and streaks which lies just below the upper chilled zone had an origin similar to the pegmatitic facies. The network of coarser-grained diabase, whose mineralogical composition is more like that of the pegmatitic facies than the finer-grained diabase of the matrix, possibly represents volatile-rich streaks in nearly solidified but still mushy magma. Because of their volatile-rich character these patches developed a coarser grain than the surrounding material. The mode of accumulation of these volatile rich streaks below the chilled roof is not entirely clear. Possibly crystallization of the roof downwards forced out volatiles into the partly liquid magma below, or bubbles of volatiles from lower down in the body moved upward and accumulated in streaks and layers beneath the solid, chilled roof.

The pegmatitic facies belongs to a late stage in the crystallization history of the diabase, possibly overlapping both the intermediate and final magmatic stages.

Granophyre.—Many writers regard the small bodies of granophyre that are associated with diabasic intrusions as having been formed by differentiation of mafic magma. Recently, however, Walker and Poldervaart (1949) in their study of the Karroo dolerites have shown that, in addition to some small bodies of granophyre thought to be of magmatic origin, there are other "metasomatic granophyres" that were formed by transfusion of sedimentary rocks by emanations from the dolerite magma. Tomlinson (1945) regards granophyric bodies he has studied at Safe Harbor, Pennsylvania, as having been generated out of granite pegmatite within the metamorphic aureole of a diabase dike.

The present writer is not prepared to state unequivocally that all the bodies of granophyre associated with diabase in Southeastern Pennsylvania crystallized from a magma, but he found no compelling field evidence in the granophyre bodies studied to indicate that they were derived from the sedimentary rocks adjacent to the diabase intrusions. The sharply terminated zircon crystals observed in the heavy residue prepared from the granophyre

in the drill hole is, however, the best criterion for the origin of the granophyre, on the premise that rounded or worn zircons could not have crystallized from a magma.

The writer regards the granophyre as the last stage differentiation product of the diabase. The residual liquid, now much depleted in volatiles and most of the mafic constituents, crystallized more rapidly to granophyre, a rock composed principally of alkali feldspar, quartz, and a little pyroxene relatively rich in iron. Thus we have an abrupt transition from relatively coarse-grained granophyric diabase to fine-grained granophyre. Quite possibly the release of pressure attending the escape of the hydrothermal solutions started a movement of residual magma under pressure from elsewhere in the diabase body toward the point(s) of lower pressure, thus augmenting the supply of dwindling magma.

The formation of granophyre from the diabase magma illustrates an important principle previously emphasized by Larsen (1938, p. 248), and Wager and Deer (1939, p. 230–231), namely, that the maximum change of composition due to differentiation of a basaltic magma is brought about only after crystallization of 75 to 80 per cent of its volume. The effect is to give an abrupt change in rock types, a contrasted suite such as we have here.

The explanation for the coarse-grained, miarolitic granophyre bodies may be that they were able to retain their hydrothermal solutions for a longer time than the fine-grained hedenbergitic granophyre. This would permit the development of larger crystals, and would also explain the predominance of hornblende over pyroxene. When the high internal pressures were relieved or overcame the external pressure, the magma began to vesiculate, but the loss of mineralizers rapidly increased the viscosity and the magma froze, trapping some of the bubbles which later yielded miarolitic cavities.

Temperature.—For reasons discussed above, it is certain that at the time of its intrusion the magma was almost entirely liquid. Sosman and Merwin (1913) determined the temperature of the liquid magma of the Palisades sill at the time of intrusion to be less than 1150°C. Hess (1941, p. 582–583) pointed out the value of pyroxenes as geologic thermometers and estimated a temperature of 1120° ± 10°C for the Palisades. Hess's method makes use of the inversion temperature of pigeonite to hypersthene. The diabase beneath the upper chilled zone at Dillsburg has pigeonite which is mostly inverted to orthopyroxene, indicating that it crystallized at or slightly above the inversion temperature, slightly more than 1100°C at this MgO:FeO ratio.

Summary

Small local bodies of granophyre occur in the upper part of diabase bodies in the Triassic (Newark-Gettysburg) basins of southeastern Pennsylvania, particularly in the Dillsburg district in northern York County. The core from a diamond-drill hole which penetrated the upper part of a large diabase body underlying the Dillsburg magnetite deposits revealed a complete section of diabase passing into granophyre.

The drill penetrated five main rock types in order: diabase chill facies, diabase, a pegmatitic facies of the diabase, granophyric diabase intermediate between diabase and granophyre, and granophyre. Granophyric diabase also underlies the granophyre and is presumably succeeded in depth by an unknown thickness of diabase not penetrated by the drill. The units grade one into the other, and there are no sharp boundaries or other features indicative of intrusive relationships.

The small lenticular masses of granophyre in diabase exposed at the surface elsewhere in southeastern Pennsylvania occur in the upper part of the diabase bodies beneath a roof of diabase. Unlike that encountered in the drill hole at Dillsburg, most of these granophyres are coarse-grained and may have miarolitic cavities. Hornblende rather than pyroxene is the usual mafic mineral.

Analyses of the chilled facies show that the original magma was of the saturated, tholeiitic type which has free quartz in the norm. By differentiation, the magma was progressively enriched in iron, alkalies, and silica. Correspondingly, the pyroxenes became richer in iron, the feldspar more sodic, and the quartz more abundant. Iron increases in the residual magma up to a late stage but declines in the

final fractionation product. Pyroxene follows the expectable trend from a magnesian variety toward iron-rich members in the late periods of crystallization, but optical data indicate that the composition lies on the diopside join in the rocks of later crystallization, and hence is rich in lime and divergent from the augite curve indicated by Hess as the expectable course in rocks of this composition.

The mechanism of differentiation at Dillsburg is visualized by the writer as follows: Subsequent to intrusion, the magma body began losing heat from the top and bottom. When the intermediate stage of crystallization was reached, a residual liquid rich in iron and titania, alkalies, and volatiles remained in the upper part of the nearly solid mass of diabase. Before complete solidification, fractures in the roof permitted the escape of volatile-rich, iron-bearing solutions which passed into the overlying sedimentary rocks, altering them and replacing layers of limestone conglomerate with magnetite. The residual liquid, now much depleted in volatiles and most of its mafic constituents, crystallized to relatively fine-grained granophyre. Where it retained its volatile constituents, large crystals developed, with hornblende predominant over pyroxene. When internal pressures reached values high enough to overcome the superincumbent load, the magma began to vesiculate; but the loss of the volatiles resulted in increased viscosity, and the magma froze, trapping some of the bubbles and developing miarolitic structure. The pegmatitic facies of the diabase commenced crystallization in the intermediate stage from local concentrations of volatile-rich magma. Because of low viscosity of the liquid, crystal growth continued over a longer period, and the minerals attained large dimensions. With continued crystallization, the remaining solution changed in composition, altering and replacing the earlier phases.

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Short Notes

STREAM PLANATION NEAR COLORADO SPRINGS, COLORADO

By J. Hoover Mackin

Introduction

Valley-widening processes in the Front Range and the Colorado Piedmont near Colorado Springs are discussed by B. A. Tator in two articles in recent numbers of the Bulletin (Tator, 1949; 1952). Because his views as to the manner of origin of terraces and benches in the Colorado Springs area differ from those based on earlier studies elsewhere along the Rocky Mountain front (*see*, for example, Lee, 1900, p. 509–511; Bradley, 1936, p. 167–169. 190; Mackin, 1937, p. 825–828) I took an opportunity afforded by a trip through Denver to visit the area to determine whether the landforms are actually different or whether the differences are matters of interpretation. The excursion was made in company with Glenn R. Scott, Richard Van Horn, and Melville R. Mudge, all of whom are now engaged in mapping bedrock and superficial deposits in other parts of the Colorado Piedmont for the Engineering Geology Branch of the U. S. Geological Survey. Their thorough knowledge of Piedmont rock types and landforms greatly extended the range and effectiveness of our observations; while we are in general agreement, the responsibility for the remarks to follow rests solely with me. These remarks are concerned with systematic geomorphology, not with the erosional history of the Colorado Piedmont.

Valley-floor Widening Processes

According to Tator, the streams of the Colorado Springs area shift laterally on their valley floors primarily by filling or blocking of the channels, and spilling into new positions, during seasonal fluctuations of discharge. Detritus resulting from weathering and erosion on the valley sides is from time to time removed from the base of the slopes by the shifting main stream; this removal permits erosion of the valley sides to proceed at the same rate from the base to the top, and the valley sides therefore retreat at a constant slope which depends on the resistance of the rock to erosion.

During certain phases of the seasonal fluctuations the stream deepens its channel through the alluvium and corrades the bedrock. Repetition of this process on every part of the valley floor, as the stream shifts from place to place by the fill-spill process, results in a tolerably smooth rock floor that is mantled by alluvium approximately equal in thickness to the depth of flood scour. The suballuvial rock surface is produced, so to speak, by innumerable saw cuts to subequal depth, with the saw held vertically.

In Tator's first paper, which describes this process of valley widening in the Front Range, lateral cutting is noted as a minor and local process in the abstract and the conclusions, but is not discussed in the body of the article. In the second paper, which applies the valley-widening theory in explaining the origin of terraces and interstream benches in the Piedmont, lateral corrasion is again considered to be an incidental process. The saw does not cut horizontally.

Tator illustrates the fill-spill mechanism of river shifting by maps showing abandoned channels, mud-flow ridges, and other details of the deposits on two narrow valley floors in the Front Range. This habit of shifting is well known (Eckis, 1928; Krumbein, 1940; 1942) to be characteristic of streams of the type described, and Tator's maps make a nice demonstration. But the fact that fill-spill is an important mechansim of stream shifting in the mountain valleys certainly does not, in itself, preclude the possibility of vigorous lateral

corrasion in bedrock by the streams, both when they are shifted against the base of the valley sides by that mechanism, and by ordinary outward cutting on bends during periods between infrequent episodes of channel choking and spilling. And whatever the relative importance of fill-spill *vs* outward cutting on bends in the sidewise shifting of streams in the mountain valleys, these do not serve, as Tator uses them, as a basis for discussion of sidewise shifting in the Piedmont, where the same streams meander on broad valley floors.

As positive evidence that the valley floors are widened by weathering back of the valley sides, and not by lateral cutting, Tator states (1) that the valley sides in a given rock type are uniform in slope, and (2) that valley floors in the Piedmont, where the rocks are weak to weathering, are much wider than those in the Range, where the rocks are resistant to weathering.

We seem to be in complete disagreement as to the facts bearing on the first point. New 20-foot contour maps of the mountain valleys in the area under discussion, and stereographic study of air photos, indicate notable inequalities in slope of the valley sides, with the steeper slopes usually on the outside of bends. In the Piedmont area which will be discussed below, the valley sides are decidedly not uniform in slope in a given rock type.

The fact that the valley floors in the weak rock of the Piedmont are wider than those in the resistant rock of the range can be considered evidence that the valley floors are formed by weathering back of the valley sides at different rates only if it is assumed that back weathering of valley sides is the only process by which valleys are widened. An obvious alternative hypothesis is that the valley floors in the Piedmont are wider than those in the range partly because the rate of lateral stream corrasion in the weak Piedmont rock is greater than in the resistant Range rock.

The accompanying stereogram (Pl. 1, fig. 1) of a pair of typical valleys in the area under discussion provides a basis for evaluating (a) outward cutting on bends *vs* the fill-spill mechanism in sidewise shifting of the stream channels, and (b) lateral corrasion *vs* weathering back of the valley sides in the widening of the present valley floors. Both of the streams show a well-developed meander pattern. Low undercut banks in the valley floor alluvium on the outside of bends, and bar and swale topography on the inside, observed in the field, indicate that the bends are shifting outward. Lateral cutting is not confined to the valley-floor alluvium; within a mile east of the highway, 15 of the bends impinge against the bedrock of the valley sides. Steep to vertical scarps cut in bedrock at these points leave no doubt as to the vigor of lateral corrasion; there is usually a sharp topographic unconformity between the raw scarp at the base and an older, soil-mantled slope developed by weathering, wash, and creep during an interval when the stream was elsewhere on the valley floor. The scarp shown in Figure 2 of Plate 1 is unusual in that it extends from the valley floor to a terrace surface; the meander at the base fits the scarp through about 40 degrees of arc.

Between the scarps now being cut, the base of the valley sides is usually concealed by talus and slope wash. Wherever there are exposures it is evident that the detritus is not a veneer on weathered bedrock, but a wedge of fill.

PLATE 1.—VIEWS OF THE LITTLE FOUNTAIN CREEK VALLEY

FIGURE 1.—STEREOGRAM OF THE VALLEY OF LITTLE FOUNTAIN CREEK AT THE MOUNTAIN FRONT (Parts of GS-ET 4-20 and 4-21, 7-27-47; *see* the Cheyenne Mountain quadrangle or Tator, 1952, Pl. 4). The main road is Colorado Highway 115. The distance from the sharp bend in the highway in the middle of the stereogram to the southeast corner is about 1¼ miles. Note the meandering habit of the streams the high meander swing scarps that form the sides of the inner valley, and the low swing and sweep scarps on the composite terrace to the north. The meander scarp shown in Figure 2 of this plate is on the north side of the valley, south of the north arrow.

FIGURE 2.—VIEW OF A PART OF ONE OF THE MEANDER SCARPS ON LITTLE FOUNTAIN CREEK The scarp is about 50 feet high, and the valley floor, to the right, is about 200 feet wide. Note the terrace gravel at the top. For location and plan view, *see* Figure 1 of this plate.

FIGURE 2

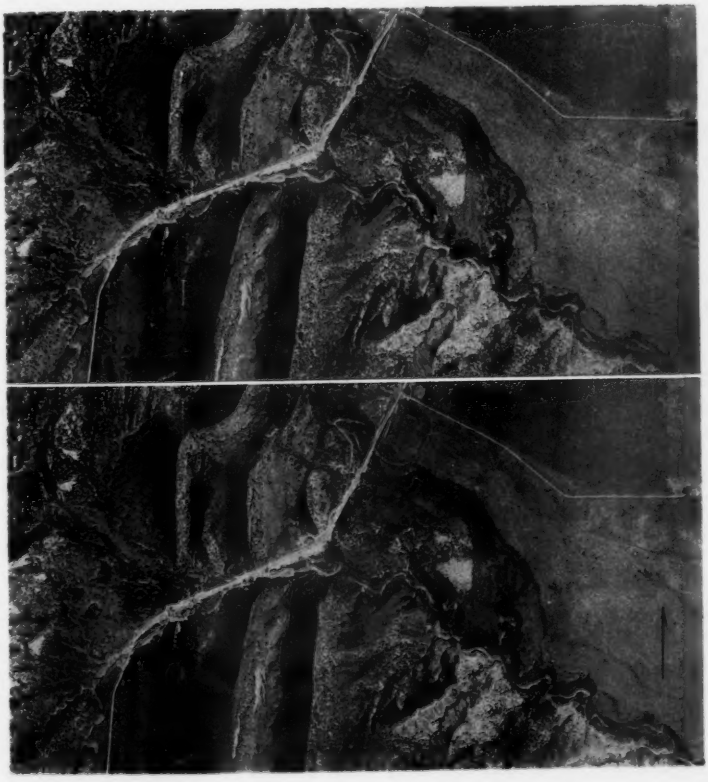

FIGURE 1

VIEWS OF THE LITTLE FOUNTAIN CREEK VALLEY

These accumulations are in a sense as significant as the fresh scarps; they indicate that there was cutting of a steeper slope by the main stream, and deposition by the valley-side modeling agencies after its withdrawal.

The valley sides of both streams east of the highway exhibit a variety of these profile forms, simple and composite, and ranging from all-scarp to all-slope. The profile forms are developed in the same weak Cretaceous shale, but there is no uniformity of slope angle; the declivities range from steep to gentle, from place to place, depending on the recency of stream cutting at the base.

In plan, the valley sides are a series of curved scarps with the same radius of curvature as the meanders. The curves meet in sharp cusps. The combination of curve and cusp is the antithesis of the plan that would be developed by slope-modeling processes with the activity of the main stream limited to removal of detritus. Figure 1 of Plate 1, viewed stereographically, leaves nothing to be said on this score.

The stereogram also shows, north of the inner valley, several terrace levels separated by meander-swing and meander-sweep scarps. The scarps between the terrace levels are similar in general to those of the inner valley sides, but they have been blunted in plan and gentled in profile in the long interval since they were cut by the stream. While bedrock exposures on the valley floor are too widely scattered to provide a picture of the detailed form of the suballuvial surface, some of the terrace scarps reveal a smoothly planed rock floor veneered by alluvium equal in thickness to the flood-scour depth of the stream channel. This is the type of surface that would be expected from the lateral corrasion now seen in operation on the valley floor. These features belong to a general group of stream-planed rock terraces, not to be confused with the equally valid group of fill terraces formed by episodes of valley filling followed by downcutting.

These several lines of evidence lead to the conclusion that the widening of the valley floors of Little Fountain and Little Bear creeks, as typical Piedmont streams heading in the Range, is due largely or wholly to lateral corrasion of the streams in bedrock, not to weathering back of the valley sides. Weathering back of the valley sides causes the slopes to lie down, not to retreat parallel with themselves. The valley-floor widening processes in the Colorado Springs area, and the interstream terraces and benches, seem to be in all essential respects identical with those described in some detail in the Big Horn Basin in Wyoming (Mackin, 1937, p. 825–828).

Reversal of Topography

Section C-D of Tator's Figure 3 (1952) shows a gravel sheet of the "lowest level of Rock Creek surfaces" passing southward beneath a higher gravel sheet labeled "middle level of Rock Creek surfaces" (Fig. 1, A). The relationship shown in the section is impossible; the younger gravel sheet of the "lowest level" cannot underlie the older gravel sheet of the "middle level."

Observations in the highway cut through which the section was drawn indicate that the gravel sheet of the "lowest level" actually terminates southward against a low, north-facing scarp, cut in rock, which represents the southern limit of river swinging at this level (Fig. 1, B). The gravel sheet of the "middle level" rests on a higher rock surface—that is, it is a remnant of a terrace that stood above the "lowest-level" valley floor. The low meander scarp that separates the two levels is now entirely buried by slope wash, the surface of which descends gently northward. The basal part of the wedge of slope wash close to the scarp contains much gravel derived from the "middle level" gravel sheet; this gravelly material grades upward into dominantly fine-textures soil with scattered pebbles, probably representing overbank silt and loess washed down in part from the "middle level" terrace, and perhaps in part from still higher uplands to the south.

This particular combination of features occurs at many places along the lateral margins of interstream benches in the Colorado Piedmont, and it is a repeating geomorphic element throughout the West wherever hard rock ranges are bordered by weak rock lowlands with gravel-veneered interstream surfaces. The mantle of wash may slope down from a higher planation

surface, as in this case, or from uplands produced by interstream erosion processes (Mackin, 1937, Pl. 4, sec. 2, 3, 4). In many instances, where the terrace or upland that was the original source has been completely destroyed, the slope wash itself, thickening toward the steep breakaway scarp at the edge of the bench, is the only evidence of the former existence of higher land flanking the bench (Mackin, 1937, Pl. 6, fig. 2; Pl. 7). This type of reversal of topography, by which a stream-planed valley floor originally bordered by uplands becomes a high-standing gravel-veneered bench on a divide, results from the great resistance of the gravel to interstream erosion processes. It is a useful concept in interpretation of divide benches because it means that each of these benches may be a remnant of an individual valley floor of limited width, rather than part of a pediment surface formerly continuous along the mountain front.

Tator's identification of Rock Creek as the stream which cut the "lowest level" (Fig. 1) raises a question which he does not discuss—namely, how did Rock Creek get from the "lowest level" surface, which is north of the "middle level" remnant, to the position where it is now incised, south of the "middle level" remnant? Some type of stream diversion seems to be required. The difficulty is that Rock Creek, a "main stream" heading in the Range, would have to be diverted by the headwaters of a very much smaller stream heading in the Piedmont; this small captor would have had none of the compensatory advantages, as difference in base level, that are commonly considered causes of capture. An explanation of why the capture occurred may lie in the common tendency for streams of small discharge, heading in weak rock lowlands, to develop graded declivities lower than those of adjacent large streams which are adjusted to the transportation of coarse-textured detritus from the mountains (Rich, 1935; Mackin, 1936; Hunt, 1946). Captures of this type, in which the cause of capture is a difference in caliber of load supplied to the contesting streams, are very frequently associated with reversals of topography

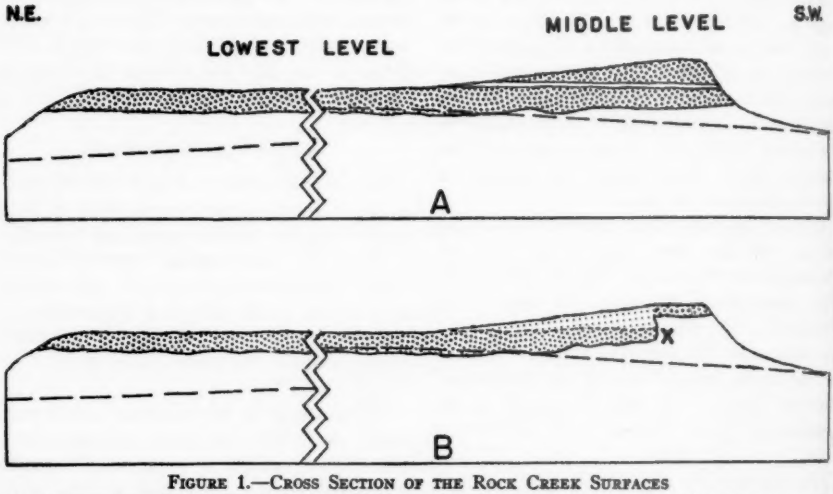

FIGURE 1.—CROSS SECTION OF THE ROCK CREEK SURFACES

A. Redrawn from Tator's Figure 3, C-D (1952). Heavy stipple is Rock Creek gravel of the Middle and Lowest surfaces, resting on Cretaceous shale. The gravel averages about 15 feet in thickness. The overall width of the bench as shown in the original is about 1400 feet. The dashed line is the profile of Colorado Highway 115. The present valley of Rock Creek is to the southwest (right-hand side of this drawing).

B. An alternative interpretation showing the Lowest-Level gravel sheet terminating southward against a meander scarp (x) cut in Cretaceous shale. The light stipple is slope wash which masks the meander scarp. The slope wash near the base of the scarp consists chiefly of gravel derived from the Middle-Level terrace.

of the type outlined above. It is interesting to note, in this connection, that Rock Creek is now, in the vicinity of the highway, about 100 feet higher than the neighboring Piedmont stream on the north.

Thickness of Alluvium on Stream-planned Surfaces

Tator notes an unusual thickness of about 100 feet of alluvium on the highest of the interstream surfaces of his Deadman Canyon group (1952, p. 261, 270). Observations along the scarp cut in this surface south of Little Fountain Creek tend to confirm his statement. In view of the foregoing discussion it is important to note that the material is "main-stream" alluvium, not slope wash derived from a former valley side.

As one of two hypotheses, Tator suggests that the thickness of the alluvium, like that on the present valley floors, may represent the depth of channel scour of the stream during the period of planation (1952, p. 270). The paper (Bailey *et al.*, 1934) cited to demonstrate that a depth of scour of this order of magnitude is not unreasonable actually deals with downcutting of streams in the Wasatch Mountains during an epicycle of erosion; it has no direct bearing on the problem of the depth of channel scour during flood stages. There seems to be good evidence that large streams, particularly those fixed in position between gorge walls, may attain depths exceeding 100 feet by scouring of movable bed materials and briefly corrading the rock floor of the channel during high-water stages. But it seems in the highest degree unlikely that depths even approaching this figure were attained by the small streams of the Colorado Piedmont, which were free to shift laterally on broad valley floors, and which could accommodate peak discharges by channel widening in alluvium as well as by channel deepening. The alluvium on lower interstream surfaces in the area under discussion averages about 10 feet and rarely exceeds 20 feet; these alluvium veneers, deposited *pari passu* with the cutting of the underlying rock surface, are the best available evidence as to the depth of channel scour by piedmont streams comparable in size with the stream that formed the Deadman Canyon surface.

As a second hypothesis Tator states:

> "The location of these excessive alluvial sections suggests that they may represent valley-profile segments where steep gradient streams encountered reduced gradients in the piedmont below the mountain base. Thicker deposits may be expectable under such circumstances" (1952, p. 270).

But thick deposits are not expectable where a steep-gradient stream encounters a "reduced" gradient in the piedmont if, as in this case, the piedmont gradient was produced in the first place by the erosional activity of the same stream operating under the same conditions. Accumulation of thick alluvial deposits represents a reversal of the normal, long-term tendency of streams to cut downward in the continental areas, and every episode of upbuilding by a stream means a change in conditions (Mackin, 1948, p. 492–499). The point made here is simply that a thickness of 100 feet of alluvium is not normal on planed surfaces cut by small streams, as implied by hypotheses 1 and 2.

Reconnaissance observations along the scarp mentioned above indicate that the thick Deadman alluvium actually consists of two depositional units, a 5- to 10-foot sheet resting on bedrock, and an overlying unit, different in texture and degree of weathering, which makes the remainder of the section. These units are separated locally by a few feet of fine-textured material similar to the overbank silt and loessic soil that mantles the gravel veneers on the other terraces and benches. The suggestion is that a period of planation was followed by an episode of aggradation. Tator mentions the possibility that there may be a relationship between the thickness of the Deadman alluvium and an episode of glaciation in the mountains (1952, p. 270). While not developed as such, this is a third hypothesis altogether different from the other two. It is to be hoped that his continuing work in the area may shed light on this problem.

Geomorphic Evidence as to Deformation

Tator concludes that there was relatively constant differential ("uparching") uplift during the time of formation of the piedmont surfaces (1952, p. 273). He rules out intermittent uplift as an explanation of the downstepping surfaces, holding that the periods of

planation correspond with arid interglacials and the periods of downcutting with humid glacials (1952, p. 267). There are two problems: (1) was there differential uplift, and (2), if so, was it constant or pulsational.

The evidence for differential movement is not clear. Some of the highest (oldest) interstream surfaces have much the same slopes as the lowest surfaces and the present streams (Tator, 1952, Fig. 2). Such differences in slope as do occur could be due (a) to crustal movement, or (b) to changes in the graded slopes of the streams. A quantitative distinction between (a) and (b) must be made if conclusions as to differential movement are to be based on a comparison of terrace profiles, but this problem is not discussed by Tator.

The definite linking of periods of planation and downcutting with interglacials and glacials, respectively, is based only on reasoning as to how the streams of the piedmont *should* respond to the climatic changes. The idea that the pauses and quickenings in downcutting indicated by terrace sequences along the Rocky Mountain front may be due wholly or in part to effects of Pleistocene climatic oscillations on the graded declivities of the streams is a useful working hypothesis (Mackin, 1937, p. 889–890; 1947, p. 113–116). The alternative hypothesis is that the pauses and quickenings in downcutting may be due wholly or in part to pulsational uplift. The two hypotheses need to be stated together to emphasize that the terrace sequence cannot be construed as evidence for either one because the other is equally competent to explain the stream behavior. The complexity of the problem is such that positive conclusions cannot be based on deductive reasoning of the type outlined by Tator.

It has been known since the turn of the century that the present relief along the Rocky Mountain front in Colorado is due primarily to differential erosion rather than to a continuation or renewal of the type of deformation that produced the range-front structures. This important generalization is based on geomorphic evidence, and the evidence is adequate to support the conclusion. But the tools available for the study of landforms will have to be better than they are now before they will be equal to the task of measuring such minor warping movement as may have occurred, steadily or pulsationally, during the erosional development of the mountain front.

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VALLEY WIDENING: A REBUTTAL

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Professor Mackin's critique has focused attention on issues that deserve careful consideration, an outstanding one of which concerns the methods of channel shifting in the modern valley flats of the Colorado Springs region. In my first descriptive work on this area (1949) I presented the thesis of valley widening by escarpment retreat and bedrock planation by vertical scour to controlled depths, the locus of the latter activity being shifted horizontally by the choking action of valley-floor alluvium. My later work (1952) developed the relationships between the modern valley flats being produced in this fashion and the higher, dissected piedmont interstream flats which occur in multiple arrangement throughout the piedmont. Mackin has presented the view that modern valley flats are being developed by gradual horizontal migration of the channeled runoff which causes escarpment retreat by undercutting and at the same time planes the bedrock of the valley floor. Thus, the perennial question is raised as to whether escarpments will retreat as a result of weathering-back without the direct aid of channeled corrasion (or sapping) along their bases.

In view of the summation of my work in the critique no further enlargement of the details is needed. The fill-spill mechanism for channel shifting, as it has been termed by Mackin, is a valid one characteristic of the stream types of the region. As stated by him, however, the recognition of this means of channel shifting does not preclude the possibility of vigorous lateral corrasion by these streams. I am in complete accord with the latter view. Nevertheless, in this particular climatic setting planation by this means is not dominant and is, as a matter of fact, limited in scope to short segments of only a few of the valleys. Evidences of lateral channel migration (lateral planation) may be observed along segments of several of the mountain-borne drainage lines. The particular stretch of Little Fountain Creek examined by Mackin (Mackin, Pl. 1) belongs in this category. However, the majority of the piedmont valleys in this area do not contain evidence to indicate that modern valley-flat development is by this means. It was not (and is not) my intent, however, to deny the efficacy of lateral planation as a process of pedimentation (valley widening) in suitable climatic circumstances.

Examination of erosion surfaces in many arid and semiarid environments, along with a close analysis of most described occurrences, has convinced me that a *continuum* of planation mechanisms exists. The particular process, or combination of processes, which dominate in an area is largely controlled by the climate. Among the climatic factors of greatest importance is the type and amount of precipitation, the latter being of course directly reflected in the runoff. Lateral planation appears to be most effective along drainage lines which are most constantly occupied by channeled runoff, as along stream lines of the "intruded" type. The latter designation includes runoff lines which gain their flow volume from adjacent more humid settings, as well as those fed by the comparatively greater precipitation (runoff) of highland areas within the dry regime. The latter is the case for a few of the mountain-borne streams of the Colorado Springs region. Fountain Creek, trunk stream of the piedmont area (1952, Pl. 1), and its major tributaries (Little Fountain, Rock, Turkey, Little Turkey, and Red creeks) gain most of their volumes from the greater precipitation of higher portions of the Front Range. Fountain Creek is able to maintain relatively constant flow throughout the year. Of the others only Little Fountain and Turkey creeks are able to maintain flowing channels in the piedmont below the mountain base, and even then only for very short distances. The stereogram and photo-

graph of Plate 1 (Mackin's critique) show evidences of lateral planation along the Little Fountain Creek segment just below the mountain base. Similar evidences may be found along the valley of Fountain creek, as well as in limited occurrences along the valleys of the streams mentioned above. Fortunately, I have been afforded the opportunity to keep these stream lines under yearly surveillance since 1940. Only during the wettest years have Little Fountain and Turkey creeks been able to maintain constantly flowing channels through the piedmont. It is along segments of these major runoff channels that evidence of undercutting of side walls has been noted. It should be clearly understood that such evidence is confined in this respect. Moreover, those valley segments containing examples of undercutting are very narrow when compared to the broader valley flats occupied by more minor piedmont drainage lines. The most extensive modern valley widening is occurring along the stream valleys which originate in the piedmont or in the lower foothills. Cognizance of the local restriction of lateral planation activity leads me to place emphasis on the shifting channel (fill-spill) mechanism.

Along those few stream segments which exhibit channel impingement against side walls (outward cutting on bends) there is also evidence of channel shifting by alluvial and other obstacles. With the exception of the channel of Fountain Creek and the limited segments of mountain-borne drainage lines noted above the streams of this region do not develop a strong meander habit. Loss of channel integrity throughout most of the piedmont (evaporation, seepage, etc.), aided by sudden shifts in channel positions during the intervals of local storm runoff, militates against the meander habit. Hence, along most of the channels lateral shift of the locus of corrasion to the side walls is more often the result of channel plugging than of gradual lateral migration (outward cutting on bends). Channel impingement against the valley walls is spasmodic and interrupted and is not the more constantly applied force usually associated with the process of lateral planation. The locus of runoff corrasion is applied only briefly, and then predominantly in the vertical sense of flood scour. The numerous abandoned channels in the valley flats, cut down to and for limited distances in the underlying bedrock, attest to the predominance of this activity. Vertical corrasion is not, however, the means for valley widening (1949, p. 1781). It is rather a bedrock-leveling process. If the valley width is to increase the side walls must retreat either by weathering-back or by the relatively constant action of water along their bases (channel or sapping action). As noted above only limited stream segments of comparatively narrow width show evidence of lateral undercutting. Moreover, there is little indication that sidewall escarpments are being steepened by sapping action. In the majority of piedmont valleys the modern channels are remote from the valley walls. Moreover, the junctions between valley sides and valley bottom are normally concealed by debris brought down by tributaries and by gravity and sheet-wash activity. In spite of these accumulations, however, the valley walls appear bold and fresh, and the bedrock lies just a few inches beneath the surface debris.

There is evidence to suggest that valley widening is being accomplished by weathering retreat of the side walls. It is on this point that Mackin's opinion differs most from mine. It is certainly undeniable that there are sharp local differences in the angles of sidewall slopes. The slope angle is in places modified by debris accumulations; in other places it is sharpened by channel impingement. Nevertheless, from a regional viewpoint the modern valley walls are strikingly uniform in the same rock type (Fig. 1). Throughout most of the piedmont area studied the valley flats and side walls are developed in the Pierre shale, a relatively weak material. The valley walls are normally capped by a layer of coarse fanglomerate, the alluvial blanket on higher, older erosion surfaces. The latter material is constantly showering down across the exposed bedrock of the valley walls. However, except for the cliff-base accumulations, the debris in transit is only inches thick. Wherever the shale is exposed it is weathered. The weathered surface material is only inches in depth, however. Debris accumulations lower down include both the fanglomeratic material and shale fragments. There is no indication that these escarpments are tending to "lie down", in spite of the general lack of recent

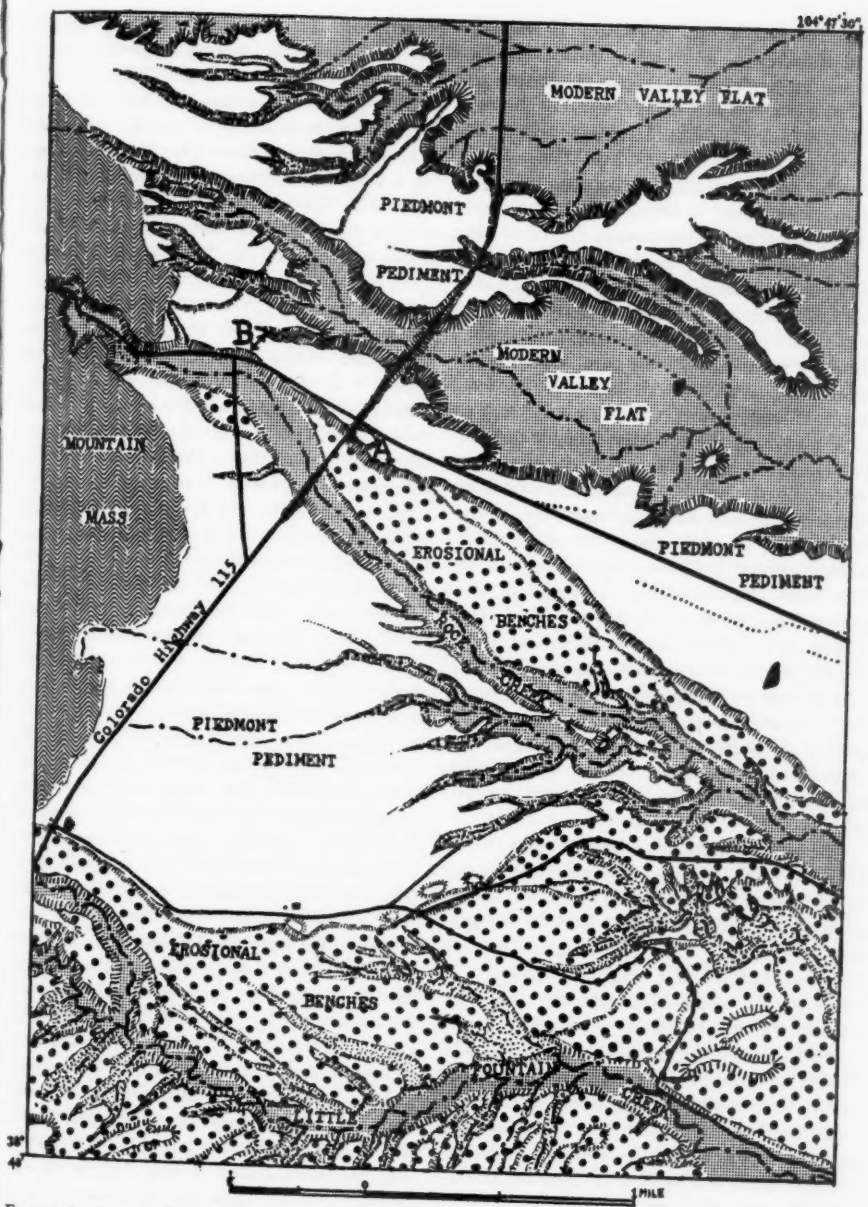

FIGURE 1.—ROUGH RELIEF SKETCH OF THE LITTLE FOUNTAIN CREEK-ROCK CREEK PORTION OF THE CHEYENNE MOUNTAIN QUADRANGLE, EL PASO COUNTY, COLORADO

The map shows the modern valley flats of the area, the erosional benches along the major stream lines, and the higher older piedmont interstream flats. "A" marks the remnant of the "middle level" of the Rock Creek group of surfaces; "B" indicates the potential future site of diversion of Rock Creek by a piedmont-borne, minor tributary of drainage north of that stream. The rough hachured effect reveals the relative uniformity of escarpments in the Pierre shale.

channel action along the cliff bases. The regional prominence of the escarpments must require a balanced interplay between rate of weathering and rate of removal of debris from the outcrop. It seems apparent that gravity and slope wash move the debris to the slope base. The mechanism for removal of basal accumulations is the action of running water, either by channel impingement or by sheet wash (or both). It is this flushing action by running water, applied during storm intervals, which allows sidewall retreat by weathering. If the removal is by main-channel impingement, and if the channel remains against the basal angle for an appropriate time, lateral undercutting will occur. However, retreat by this means is a small portion of the total for valley widening. The locus of such corrasion activity is applied but briefly during the short storm intervals. Moreover, alluvial damming tends eventually to shift the channel position to some other locus of activity. The interlacing pattern of abandoned channels found in many of the piedmont valley segments is evidence of this. A constantly applied locus of lateral corrasion is quite impossible under the influences of intermittent runoff which is responsible for the rapid shifts of channel positions. Only along those stream segments where channel integrity is maintained by constant runoff can lateral planation be the predominant process of valley widening.

The error detected by Mackin (Tator, 1952, Secion C-D, Fig. 3) is an inexcusable oversight made in drafting the figure. The "lowest level" north of the modern valley of Rock Creek does terminate against a low, north-facing escarpment as indicated in the figure. This scarp is obscured by slope wash derived from the slightly higher "middle level". I know of no evidence to favor interpretation B of Mackin's Figure 1, although this possibility should certainly be considered. The masking of older escarpments by slope wash from higher areas is common in this region. There is no longer an efficient means for removal of these accumulations from the older escarpments.

There is no direct reference in my writings (1952, p. 262) which identifies Rock Creek as the stream responsible for development of the Rock Creek Surfaces. The latter term was chosen to designate the group of related erosion levels associated in geographic position with that stream. Remnants of the "highest level" of the group could not possibly have been cut by that stream inasmuch as upslope continuations of the level place the apical portion "in a westward indentation of the foothills at Lime Kiln Creek" (1952, p. 262). Similarly, there is little certainty that all of the "middle level" remnants of the Rock Creek Surfaces north of Rock Creek were cut by that stream. On the other hand, there is petrographic evidence that Rock Creek had an early course across the "middle level" remnant shown in Section C-D of Figure 3 (1952). Certain constituents of the alluvium of this remnant appear to be erosional derivatives of diagnostic pegmatite outcrops occurring in the mountain canyon segment of Rock Creek. It is my opinion, however, that the large "low-level" remnant which forms the modern divide just north of Rock Creek was developed by that stream. Two possibilities may provide explanation for the "middle-level" remnant between the "low-level" valley flat and the present valley of Rock Creek. (1) The mountainward portion of the present "low-level" remnant may have been developed by a piedmont-borne tributary of Rock Creek, the former mountainward extent of the "middle level" divide having been destroyed by the northward recession of the north valley wall of the modern valley of Rock Creek. Similar broad tributary flats, which merge downstream with the valleys of mountain-borne drainage, occur throughout the piedmont. (2) The more likely possibility, however (as suggested by Mackin), involves drainage diversion of a type not uncommon in the piedmont. One such potential diversion has been noted by me (1952, p. 271; point C, Fig. 4). As stated, "Diversions of this sort may be responsible, at least in part, for local variations in gradient, slope direction, and dimensions of the piedmont interstream surfaces". Rock Creek may have swung through a slightly more northerly course during the "low-level" interval (Pl. 1). Diversion by headward extension of a more southerly drainage line may have developed the present valley position. Mackin notes that the modern valley of Rock Creek is at a greater elevation just below the

mountain base than the floor of the neighboring piedmont stream to the north. Headward extension and valley widening along the latter (Pl. 1) would create a circumstance similar to that of the tributary of Little Bear Creek with Little Turkey Creek as noted by me (1952, p. 271–272). If such were to occur, the present valley flat of Rock Creek would be left separated from its new course by a sizable remnant of the older surface (low level of the Rock Creek Surfaces).

The excessive alluvial cover (100 feet or more) on the highest interstream surface of the Deadman Canyon group (1952, p. 270) presents a problem. If the alluvial cover were contemporaneous with the interval of planation it would have required, as stated by me, "... a depth of scour greater than that characteristic of the modern valley bottoms." I observed further that a depth of scour of 100 feet is not unreasonable and cited Bailey's work (1934) on floods and accelerated erosion in Utah. As noted by Mackin, the latter work deals with an epicycle of erosion during which cumulative scour by flood waters produced deep channels in excessive alluvial deposits. It was in reference to such cumulative scour activity that I cited several examples in my "valley-widening" paper (1949, p. 1781) and observed that "... in other regions affected by torrential runoff, scour activity reaches to considerable depth in the valley bottom alluvium." A repetition of channel deepening along successive channel positions should provide the same mechanism for bedrock corrasion as the single flood in thinner alluvium. Moreover, corrasion to bedwork through considerable thickness of alluvium is not necessarily restricted to the large stream channels. Troxell and Peterson (1934) describe flood scour of this nature along the La Canada Valley, California, a setting comparable to that discussed herein. The mountain-borne streams of La Canada Valley are intermittent, and each has built a debris cone on the valley floor at the canyon mouth. The authors state (1934, p. 73), "During periods of small storm runoff the streams reach the top of the cone with little or no debris load ... pick up debris ... gradu-al(ly) deepening" the channels across the cone. Channels in these cones up to 150 feet deep often reach the underlying granite of the main valley floor. On the other hand, during periods of more intense runoff, old cones may be obliterated and new cones developed. It necessarily follows that new channel positions will be created. Repetition of the process of small-storm deepening and large-storm shifting of channels may be a feasible means for rock-floor planation beneath excessive valley fill. It is not my intent here to convey a personal conviction that the erosion level in question was corraded under its present thick alluvial blanket. On the contrary, it is my opinion that excessive alluvium on pediment surfaces is generally the result of post-pedimentation accumulation.

I do not think it can be denied that the piedmont alluvial fan at the mouth of a mountain canyon is a thicker accumulation of debris "where steep-gradient streams encountered reduced gradients ..." (1952, p. 270). "The most abrupt gradient changes" in the Colorado Springs region "occur at the mountain base" (1952, p. 259). This appears to be the case for the thick deposit in the highest level of the Deadman Canyon surfaces. The basal 5- to 10-foot unit of this section mentioned by Mackin may represent the flood-plain deposit contemporaneous with the development of the underlying surface of planation. However, in my studies over a period of years I was unable to establish any considerable lateral continuity for this unit, although it is certainly recognizable locally. Nevertheless, the presence of a broad, thinly alluviated, pre-aggradational piedmont flat seems to be indicated. It should be noted that the alluvium of this remnant thins upslope and downslope from the maximum noted. Although other fragments of this level are preserved in this region none is as extensive in longitudinal profile continuity as this one. Thus, similar alluvial thicknesses may have been present on other ancient piedmont flats of the same age. If, following an interval of valley widening in the piedmont, there occurred an episode of increased precipitation in the mountains, it seems likely that alluvial fans would be built outward on the valley floors at the mouths of the mountain canyons. This activity is common in dry regions where localized storms of torrential nature, able to

move large quantities of debris through their steeper mountain canyon segments, are nevertheless incapable of moving this debris any distance outward from the mountain base. Such an aggradational phase might encompass a relatively short time interval, perhaps associated with only a slight shift in precipitation conditions. On the other hand, there exists the as yet unexplored possibility (in this particular region) that there may have been some relationship between aggradation of these old valleys and glaciation in the mountain area, as I suggested (1952, p. 270). However, this is as yet only conjecture.

My conclusions concerning the regional relationships of these erosion levels are certainly based more on deductive reasoning than on quantitative evidence. However, in the absence of sharp distinctions as to regional controls I feel there is justice in the deductive approach. Although I agree with Mackin that explanation of the regional multiplicity of these levels cannot be positively ascertained to be the result of climatic controls, the weight of qualitative "evidence" points in that direction. There is little difference between the profiles of the modern valley flats and the highest of the interstream surfaces (1952, Fig. 2). On the other hand, in view of the location and relatively short longitudinal extents (eastward) of these levels, profile variations due to intermittency of mountain uplift would not be expectable. There is no evidence known to me of structural warping along the Rocky Mountain front during the late Quaternary. Any mountain uplift which may have occurred has surely also extended its influence far to the east of the downslope extent of the levels under discussion here. I failed in my writings to make a distinction between differences of slope due to crustal movement and those due to stream-grade changes simply because detectable slope differences of this nature do not exist.

The tentative hypothesis presented by me (1952, p. 273) that slight climatic alternations between semiarid and subhumid may account for the differentiation of the piedmont levels was based largely on the generally accepted fact that climatic changes have occurred throughout the Pleistocene. Moreover, the important generalization that the modern Rocky Mountain front landscape is due primarily to differential erosion is pertinent in this respect. The multiplicity of these levels is also regional in scope. It seems far more rational to place regional landscape elements under the category of climatic control than under that of minor structural warping.

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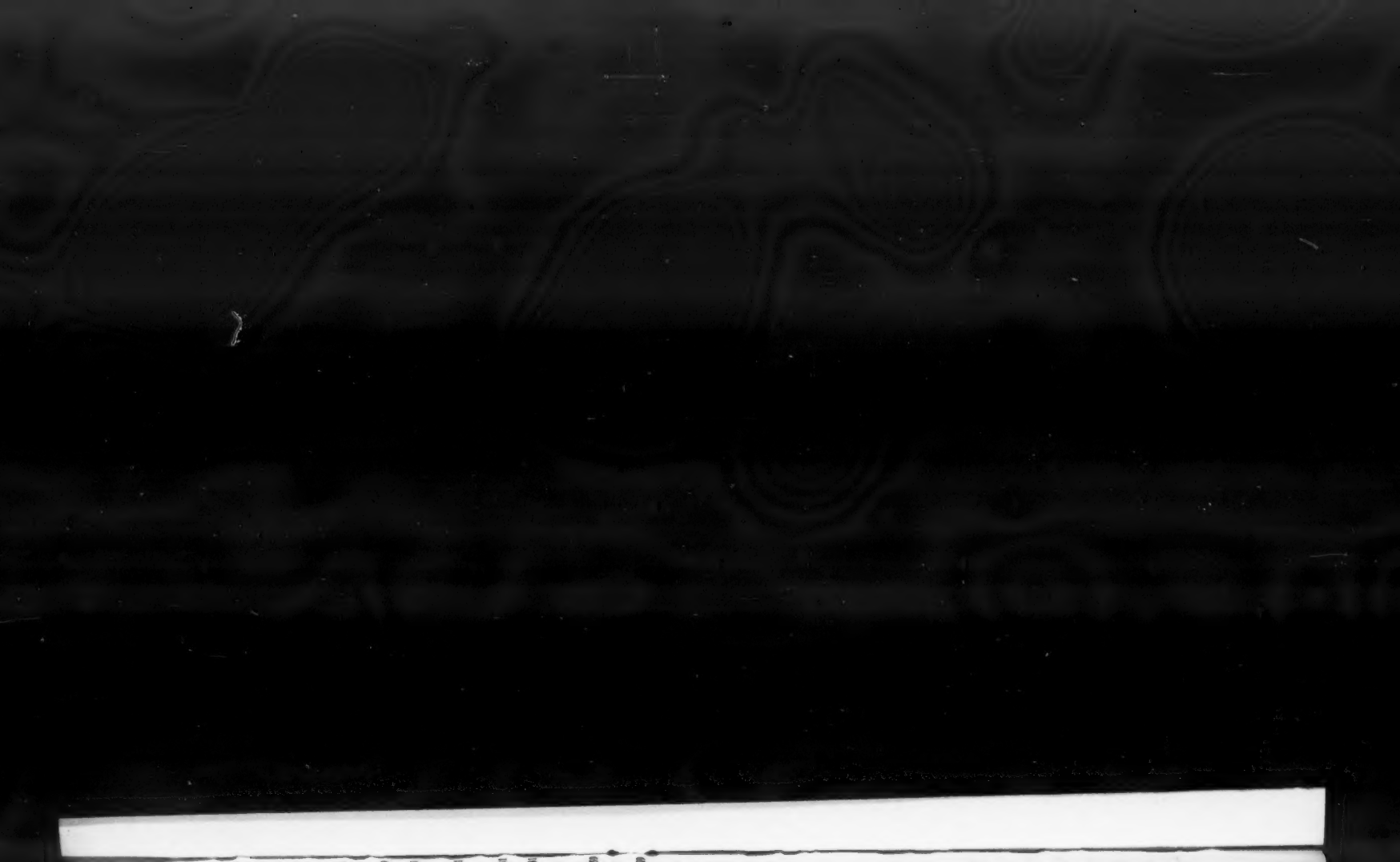
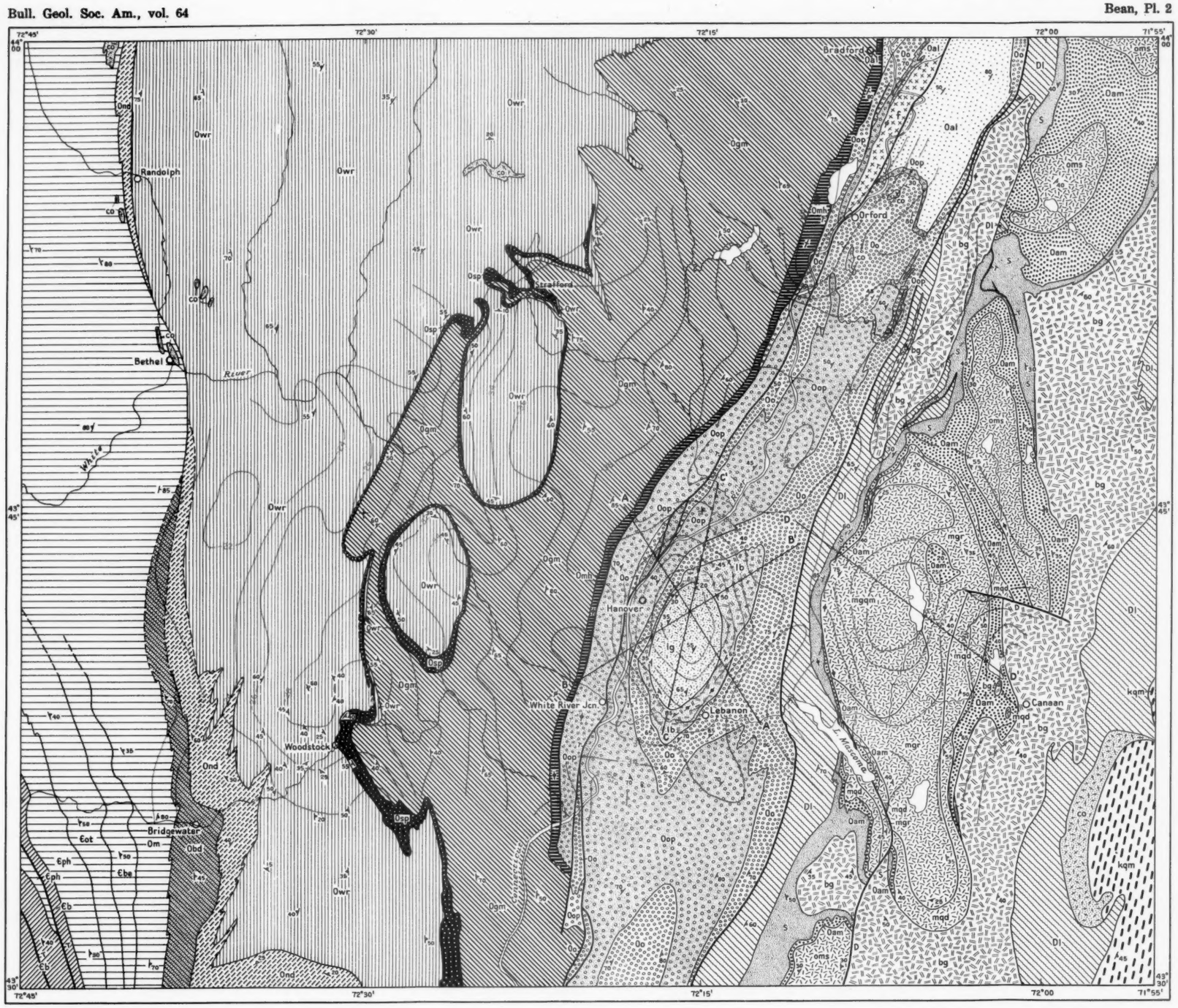
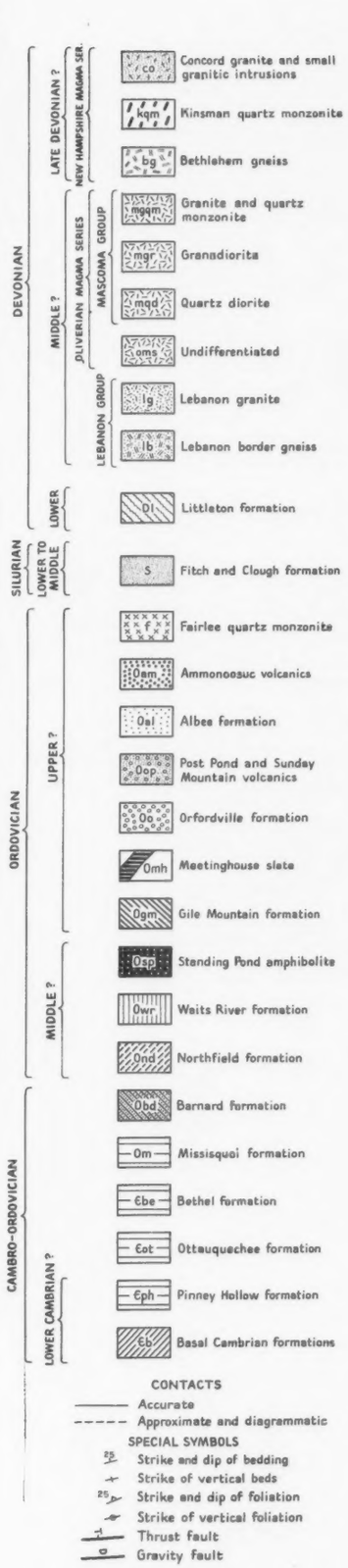
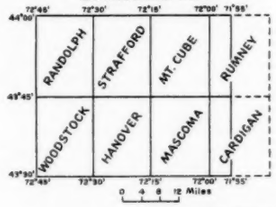






QUADRANGLE INDEX



Scale

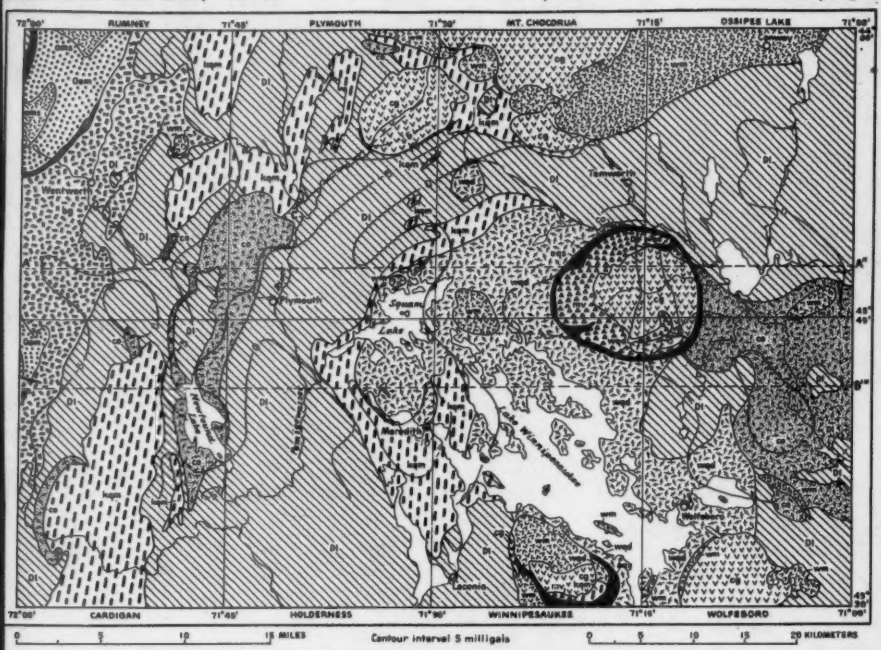
1 0 1 2 3 4 5 6 7 Miles

1 0 1 2 3 4 5 6 7 Kilometers

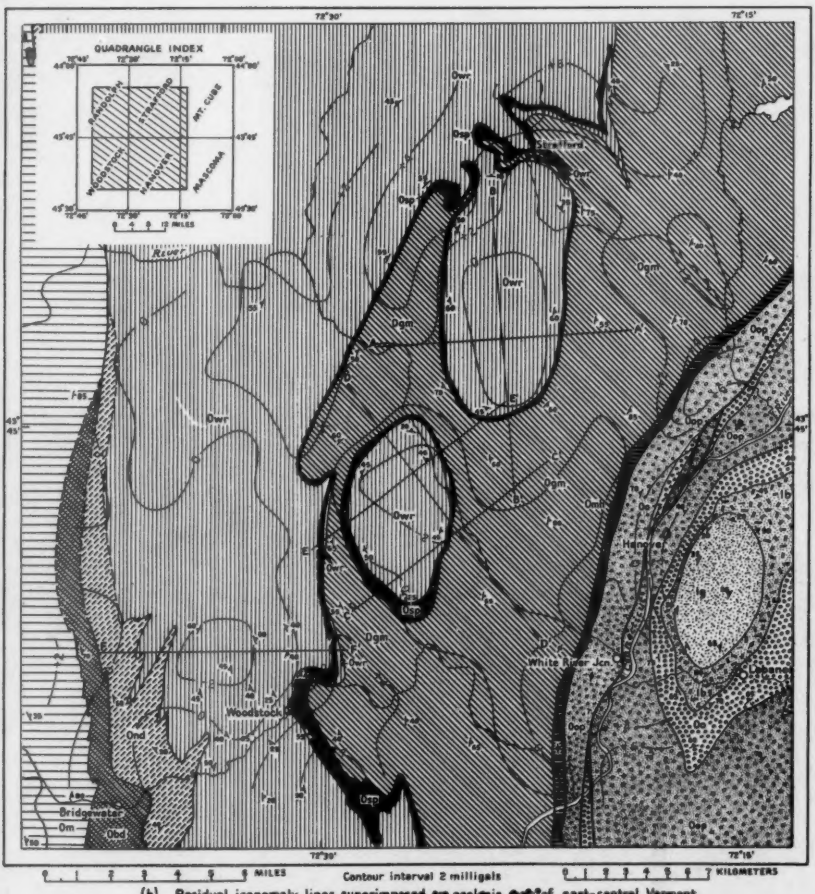
Contour interval 2 milligals

COMPLETE BOUGUER ISANOMALY LINES SUPERIMPOSED ON GEOLOGIC MAP OF EAST-CENTRAL VERMONT AND WEST-CENTRAL NEW HAMPSHIRE

DET·PUB·LIB



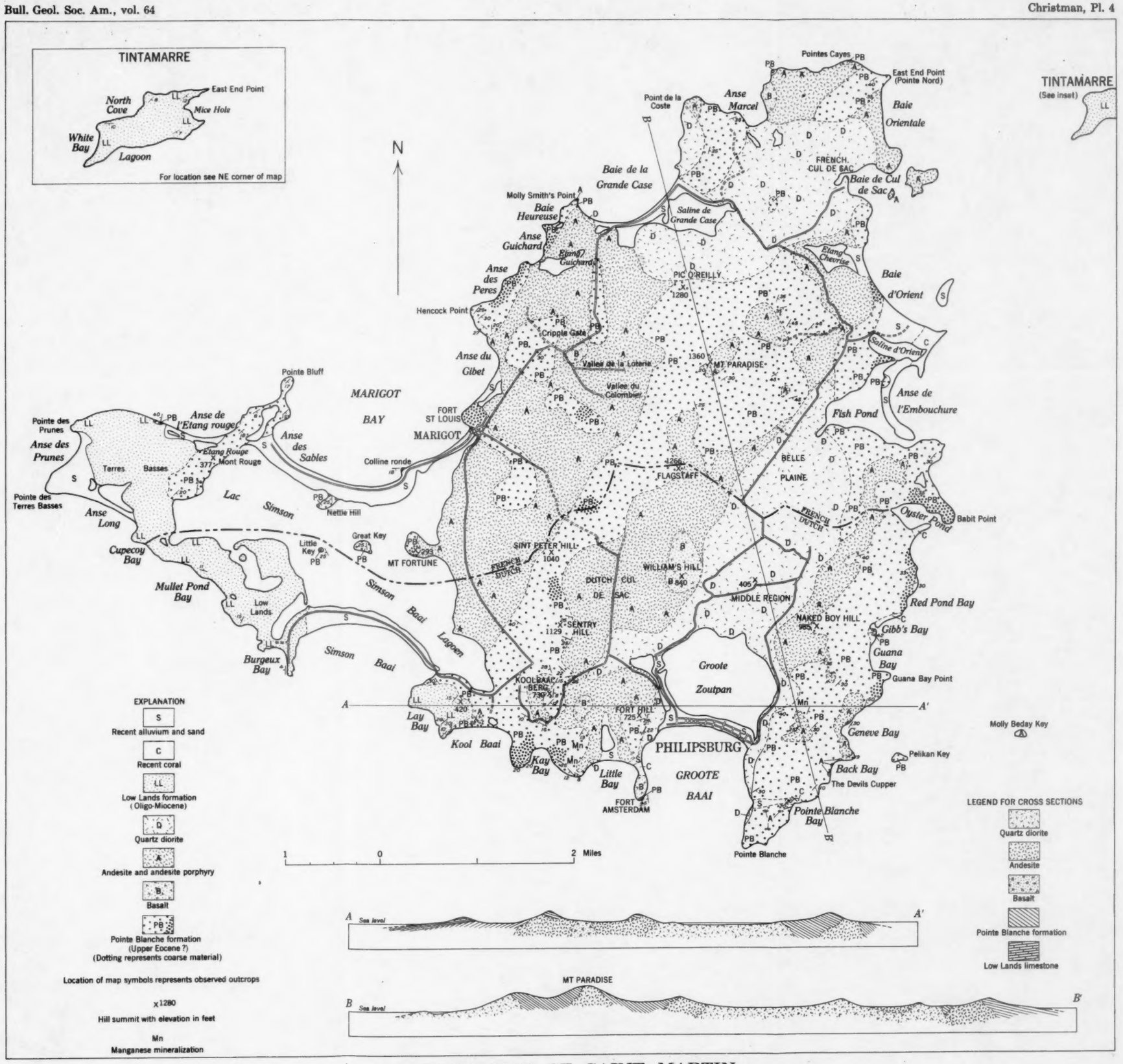
(a) Residual isanomaly lines superimposed on geologic map of east-central New Hampshire



(b) Residual isanomaly lines superimposed on geologic map of east-central Vermont.

RESIDUAL ANOMALY MAPS

Refer to Plate 1 for legend



GEOLOGIC MAP OF SAINT MARTIN

DET·PUB·LIB

ISOPACH MAP OF CENOZOIC BASIN FILL IN THE DELAWARE BASIN, TEXAS AND NEW MEXICO

Contour Interval - 100 feet

Capitan reef front

Well location

Isolated bedrock outcrops

0 5 10 15 miles

HOBBS

Area of scattered outcrops

LEA

GAINES

CARLSBAD

Pecos River

EUNICE

A'

NEW MEXICO

TEXAS

ANDREWS

Area of very thin fill

OTERO

EDDY

JAL

NEW MEXICO

TEXAS

LOVING

KERMIT

WINKLER

A

B'

CULBERSON

MONAHANS

Limit of Fill

REEVES

B

BARSTOW

WARD

PECOS

Western

Approximate

TOYAH

C'

C

VAN HORN

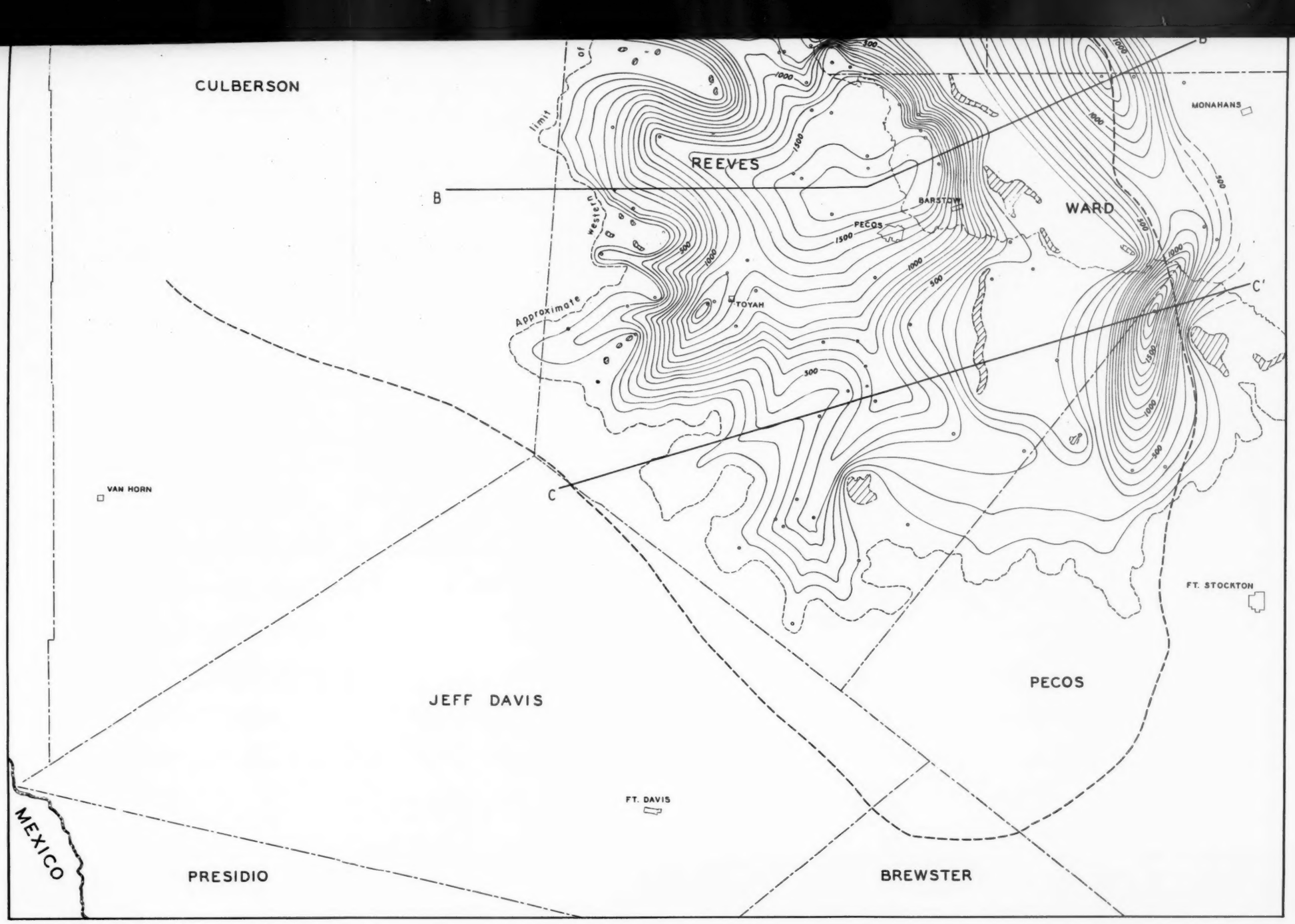
FT. STOCKTON

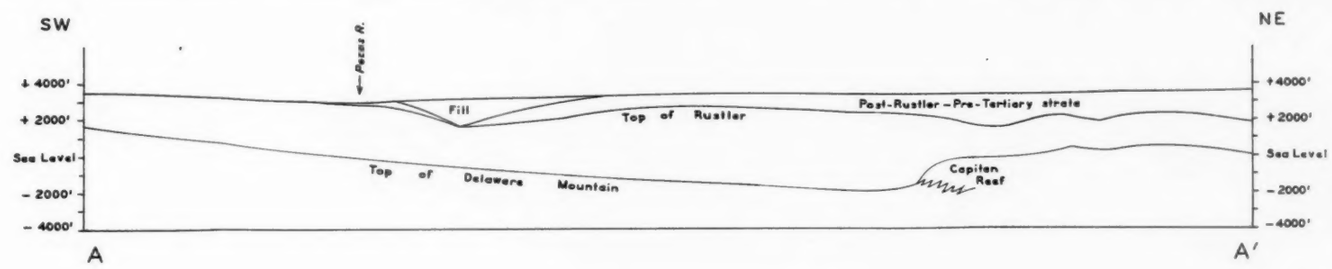
PECOS

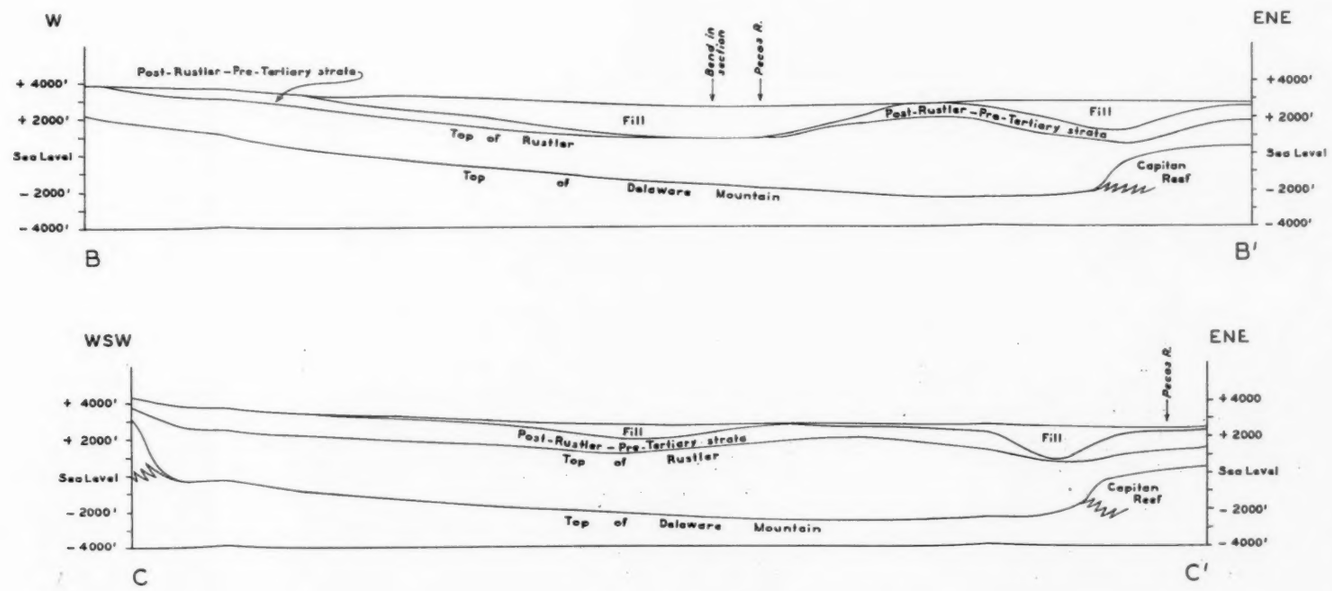
JEFF DAVIS

FT. DAVIS

MEXICO







ISOPACH MAP OF CENOZOIC BASIN FILL AND CROSS SECTIONS

DET·PUB·LIB

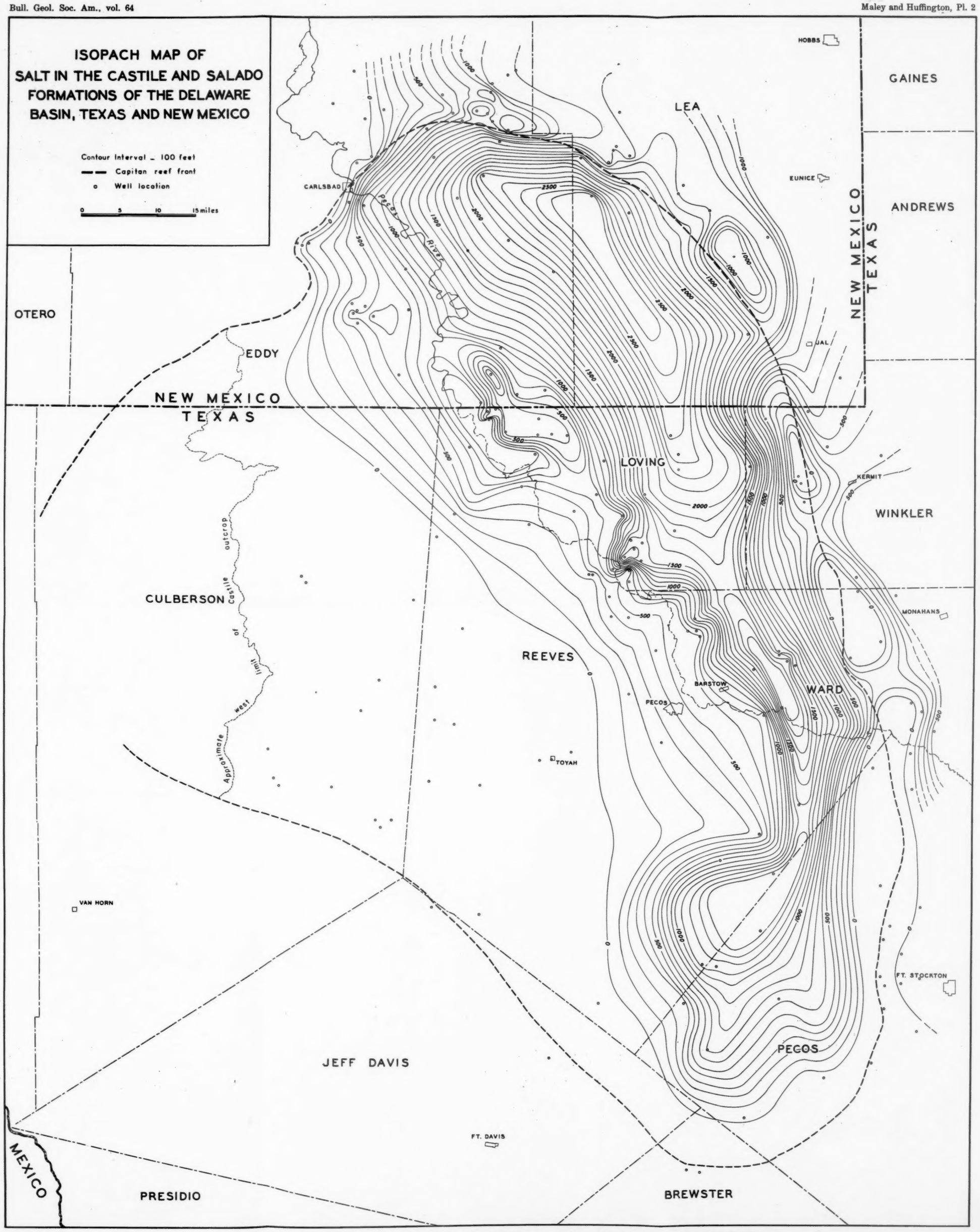
ISOPACH MAP OF SALT IN THE CASTILE AND SALADO FORMATIONS OF THE DELAWARE BASIN, TEXAS AND NEW MEXICO

Contour Interval – 100 feet

Capitan reef front

o Well location

0 5 10 15 miles



ISOPACH MAP OF SALT IN THE CASTILE AND SALADO FORMATIONS

DET·PUB·LIB

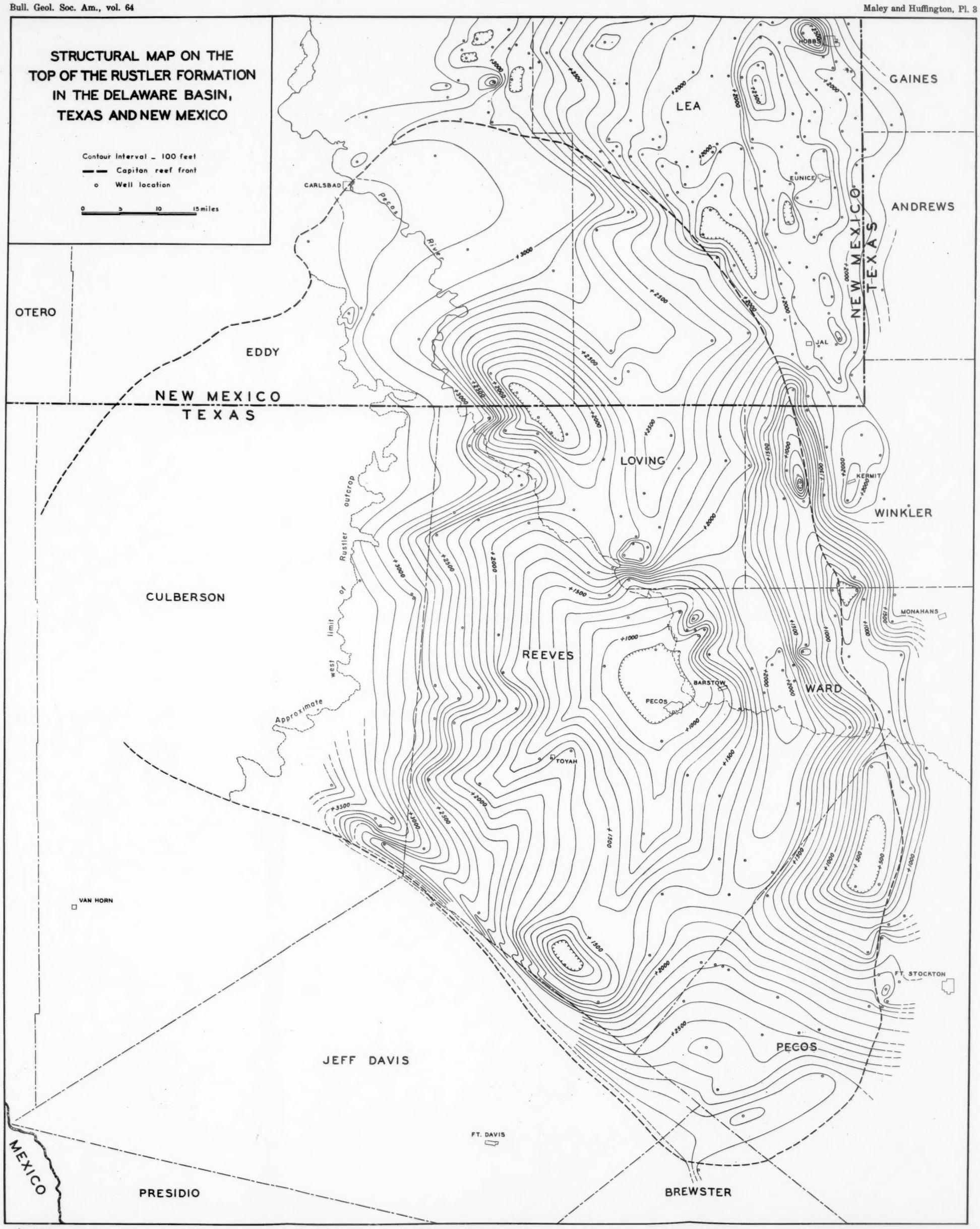
STRUCTURAL MAP ON THE TOP OF THE RUSTLER FORMATION IN THE DELAWARE BASIN, TEXAS AND NEW MEXICO

Contour Interval – 100 feet

Capitan reef front

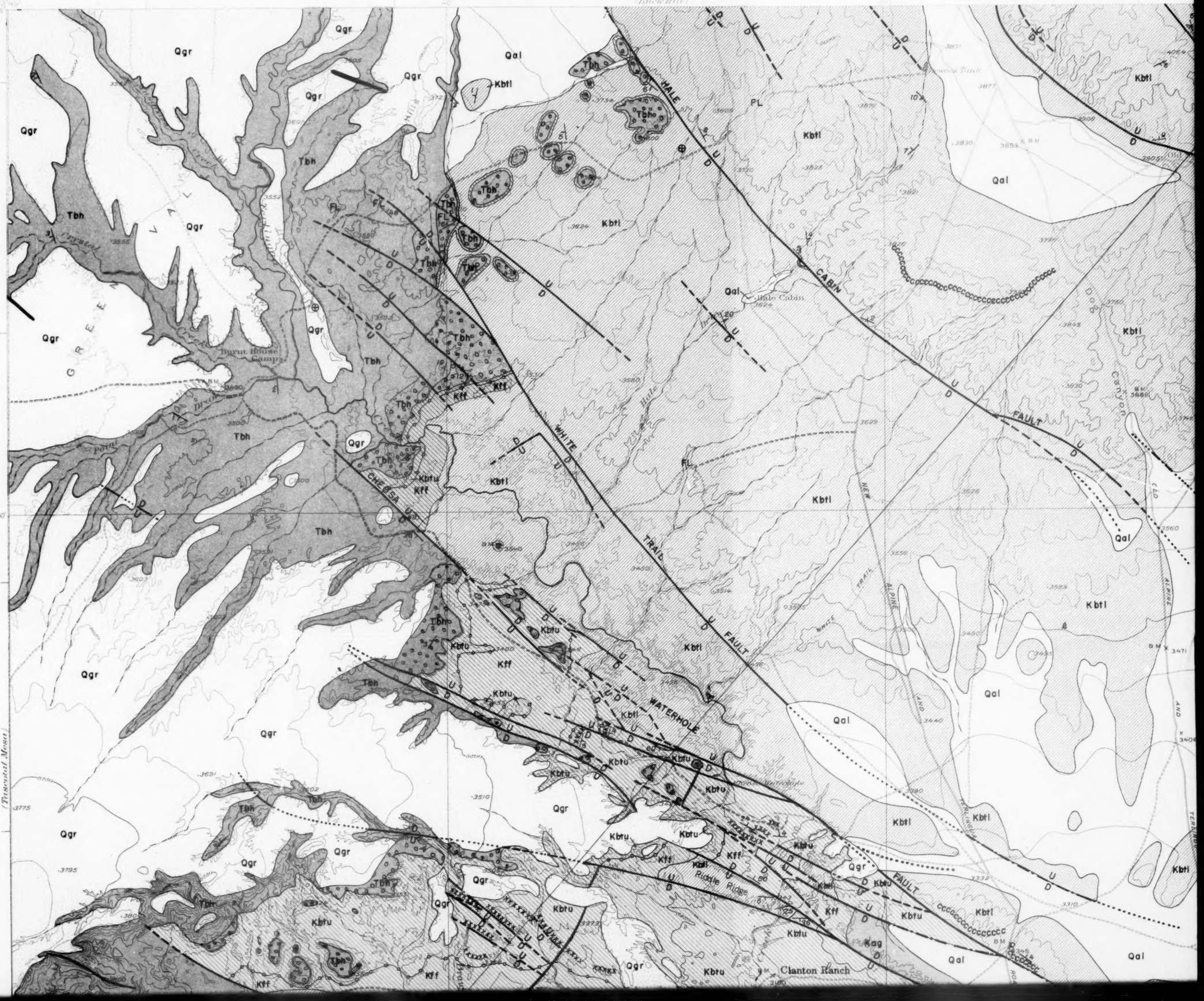
Well location

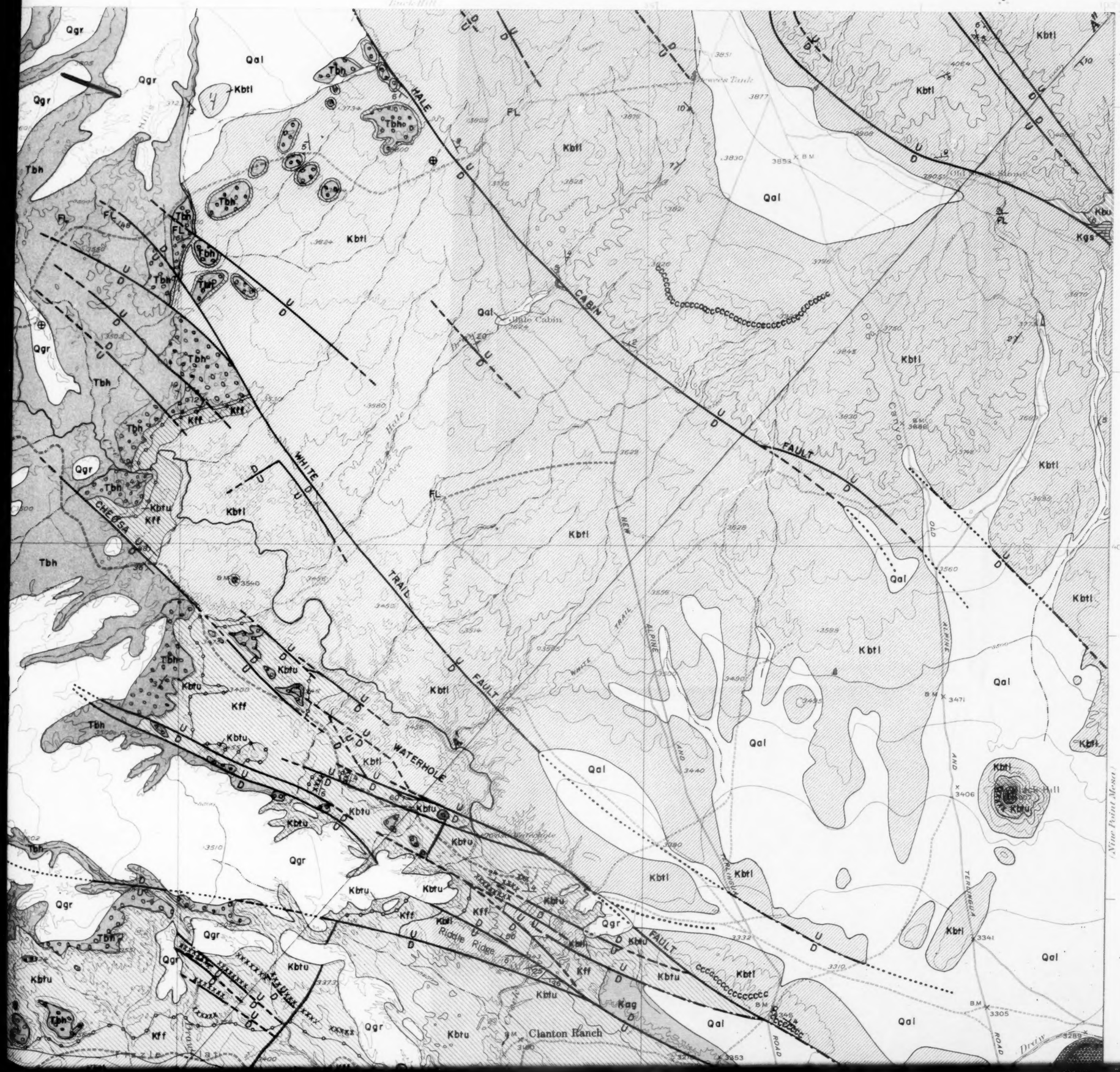
0 5 10 15 miles



DET·PUB·LIB

STRUCTURAL MAP ON THE TOP OF THE RUSTLER FORMATION


Qgr
Qal
Kbtl
Tbh
Kff
Kbtu
Kag
HALE
CABIN
FAULT
WHITE
TRAIL
CHEESA
WATERHOLE
Ridge
Clanton Ranch
Burnt House Camp
Hale Cabin
Crystal
Portal
Trail
Alpine
Pinnacles
Buck Hill
Hills
Creek
Canyon
Dry
Tunnel Mtn.
GREEN VALLEY



EXPLANATION

QUATERNARY

Qal — Alluvium

Qgr — Terrace gravels

TERTIARY

IGNEOUS ROCKS

Intrusives

Lava flows

Tbh — Buck Hill volcanic series including basal conglomerate.

UNCONFORMITY

Eocene? – Oligocene?

Kag — Aguja sandstone

CRETACEOUS

Gulf series

Kbtu — Upper unit (approximate equivalent of Udden's Terlingua beds) including Fizzle Flat lentil at base.
xxxxxxx: *Inoceramus undulato-plicatus* zone.

Boquillas – Terlingua

Kbtl — Lower unit (approximate equivalent of Udden's Boquillas flags)
ccccc: *"Crioceras"* zone.

UNCONFORMITY

Comanche series

Kbu — Buda limestone

Kgs — Grayson marl

Kdv — Devils River limestone

Normal Faults

U / D

U, upthrown side.
D, downthrown side.

Inferred

Downthrown side covered by alluvium; position marked by scarp.

DEVIL'S GRAVEYARD

TASCOTAL MESA

SOLITARIO

WATERHOLE FAULT

FIZZLE FLAT FAULT

Riddle Ridge

Clanton Ranch

Whistler Squat

Half Dome

Gray Hill

Tbh Tertiary basalt and andesite

Qgr Quaternary gravel

Qal Quaternary alluvium

Tf Tertiary felsite

Ti Tertiary intrusive rocks

Kag Agua Formation

Kbtu Upper Bissett Member

Kbtl Lower Bissett Member

Kff Fort Field Member

Kdv Del Valle Member

Kgs Gray Sandstone Member

Kbu Buda Limestone

Contour interval 50 feet.
Datum is mean sea level.

Geology by C.

Section A-A'

Comanche series

Buda limestone

Kgs

Grayson marl

Kdv

Devils River limestone

Normal Faults

U / D

U, upthrown side.
D, downthrown side.

Inferred

Downthrown side covered by alluvium; position marked by scarp.

Concealed by alluvium.

Doubtful fault

Strike and dip of beds

Beds essentially flat lying

FL

Fossil locality

Clanton Ranch

Riddle Ridge

Half Dome

Gray Hill

Camels Hump

Geology by C. G. Moon, 1947 and 1948

(Tertiqua 125000)

Scale 1/25000

1 0 1 2 3 4 Miles

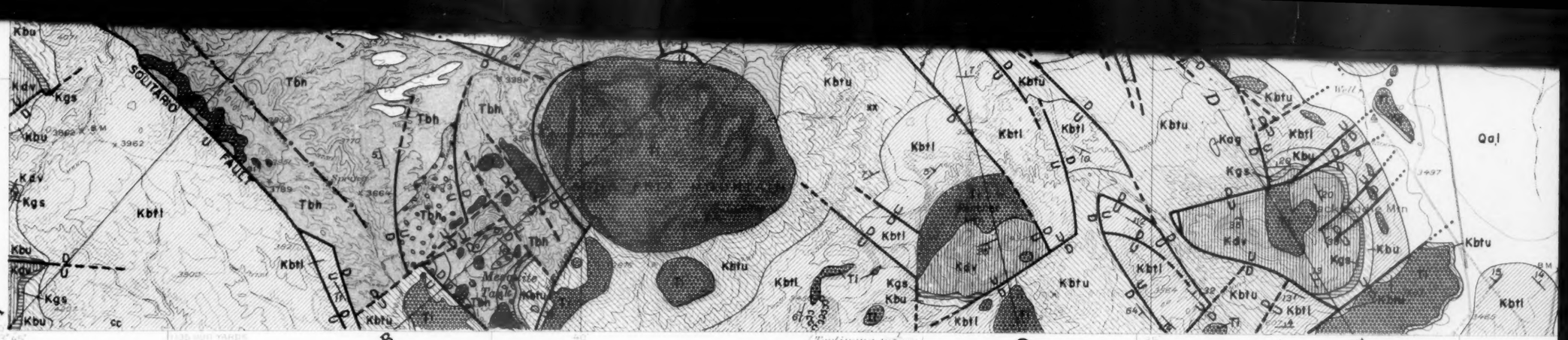
5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 5 Kilometers

Contour interval 50 feet.
Datum is mean sea level.

APPROXIMATE MEAN DECLINATION, 1947

A' 4000 3000 2000



(Tertiary) 125mm

Scale 1/62500

1 0 1 2 3 4 Miles

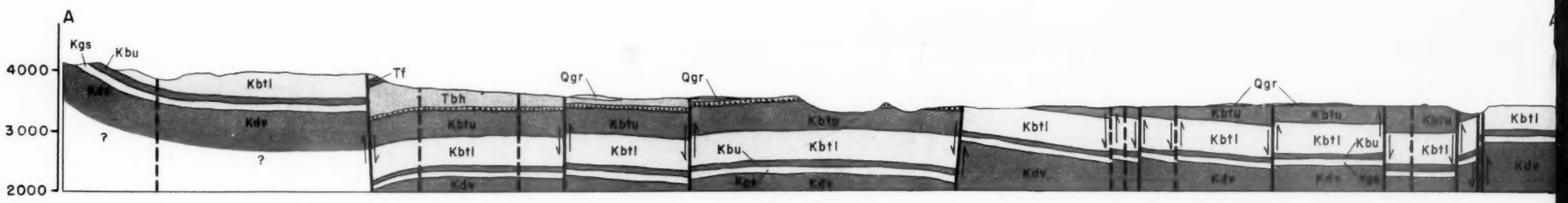
5000 0 5000 10000 15000 20000 Feet

1 0 1 2 3 4 5 Kilometers

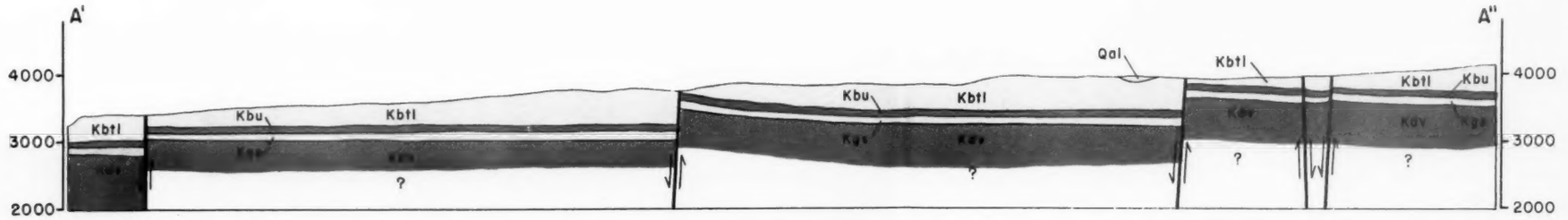
Contour interval 50 Feet

Datum is mean sea level.

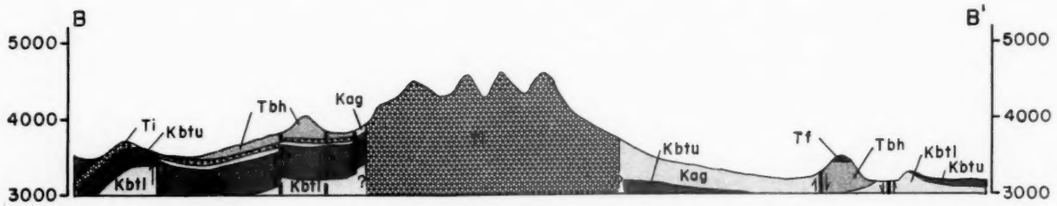
Geology by C.



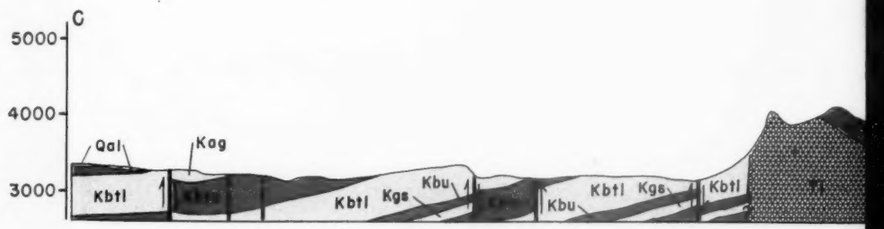
Section A-A'



Section A'-A"

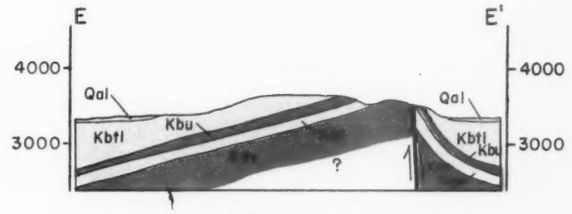


Section B-B'



Section C-C'

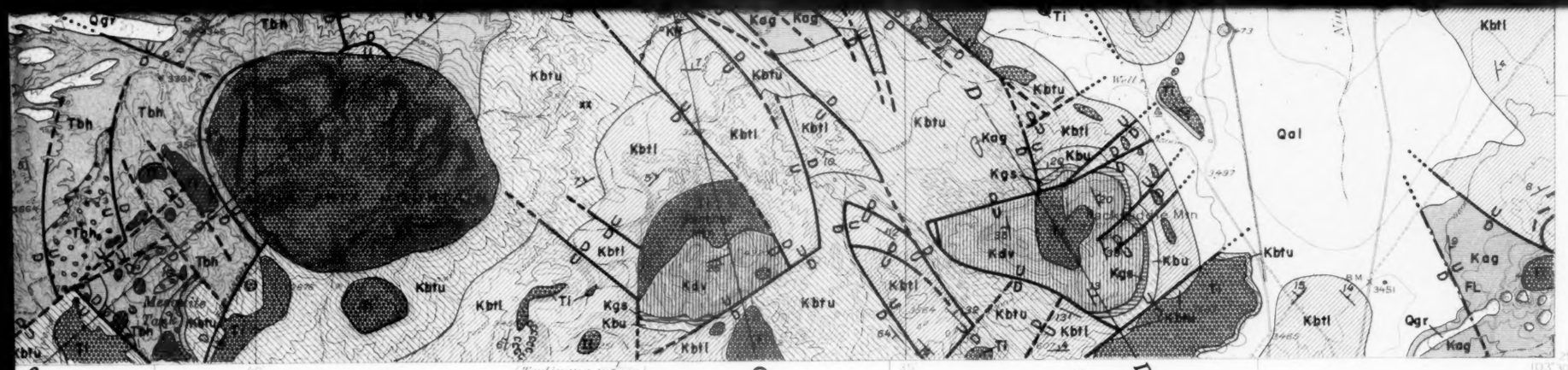
Section D-D'



Section E-E'

GEOLOGIC MAP AND STRUCTURE SECTIONS OF AGUA FRIA QUADRANGLE, T

DET-PUB-LIB



(Tertiapat 12500)

Scale 1:2500

1/2 0 1 2 3 Miles

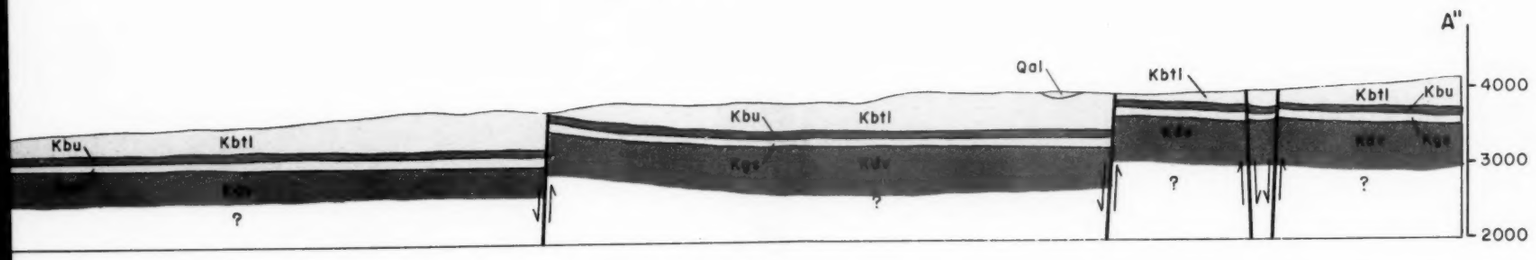
5000 0 5000 10000 15000 20000 Feet

1/2 0 1 2 3 4 5 Kilometers

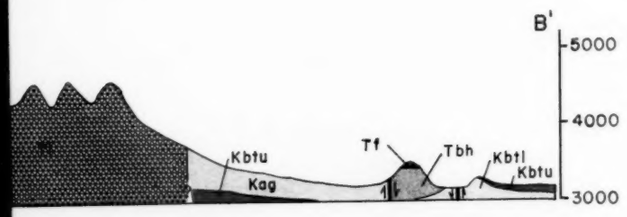
Contour interval 50 feet.
Datum is mean sea level.

Geology by C. G. Moon, 1947 and 1948

Section A-A'



Section A'-A''



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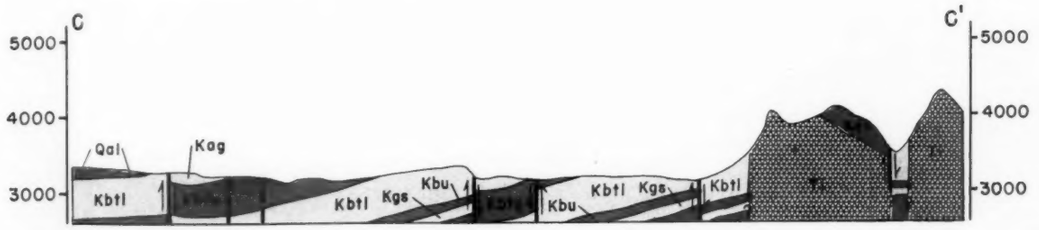

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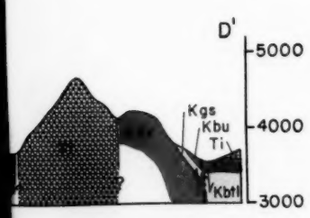
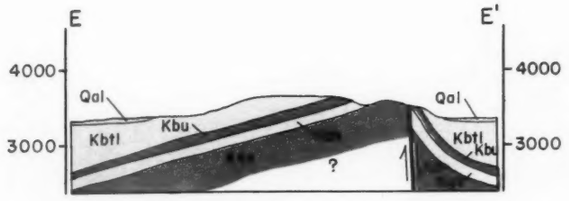


ction D-D'



Section E-E'

AND STRUCTURE SECTIONS OF AGUA FRIA QUADRANGLE, TEXAS

67°10' 67°05' 67°00' 66°55'

10° 35'

10° 30'

MAR CARIBE

A

B

CABO BLANCO

CATIA DE LA MAR

MAMO

MACUTO

MAIQUETIA

LA GUAIRA

LAS PALMAS

MACUTO

AGUA

PEÑA DE MORA

LOS CASTILLOS

INFIERNITO

EL MANICOMIO

CARACAS

PRO-PATRIA

EL PARAISO

EL CEMENTERIO

LA VEGA

ANTIMANO

ELVALLE

TUNQUITO

ANTICLINE

Río Guaire

Río Valle

Río Macarao

LAS ADJUNTAS

LAS BRISAS

67° 00'

66° 55'

66° 50'

66° 45'

C A R I B E

A

CABO BLANCO

B

MAQUETIA

LA GUAIRA

C

MACUTO

10° 35'

PEÑA DE MORA

LOS CASTILLOS

INFIERNITO

PICO DE NAIGUATA

LA SILLA

EL MANICOMIO

CARACAS

LA FLORIDA

PRO-PATRIA

10° 30'

EL PARAISO

CIUDAD UNIVERSITARIA

CHACAO

DOS CAMINOS

Rio Guaire

EL CEMENTERIO

LA VEGA

PETARE

ELVALLE

ANTIMANO

Rio Valle

EL ENCANTADO

LA GUAIRITA

LIRA

BARUTA

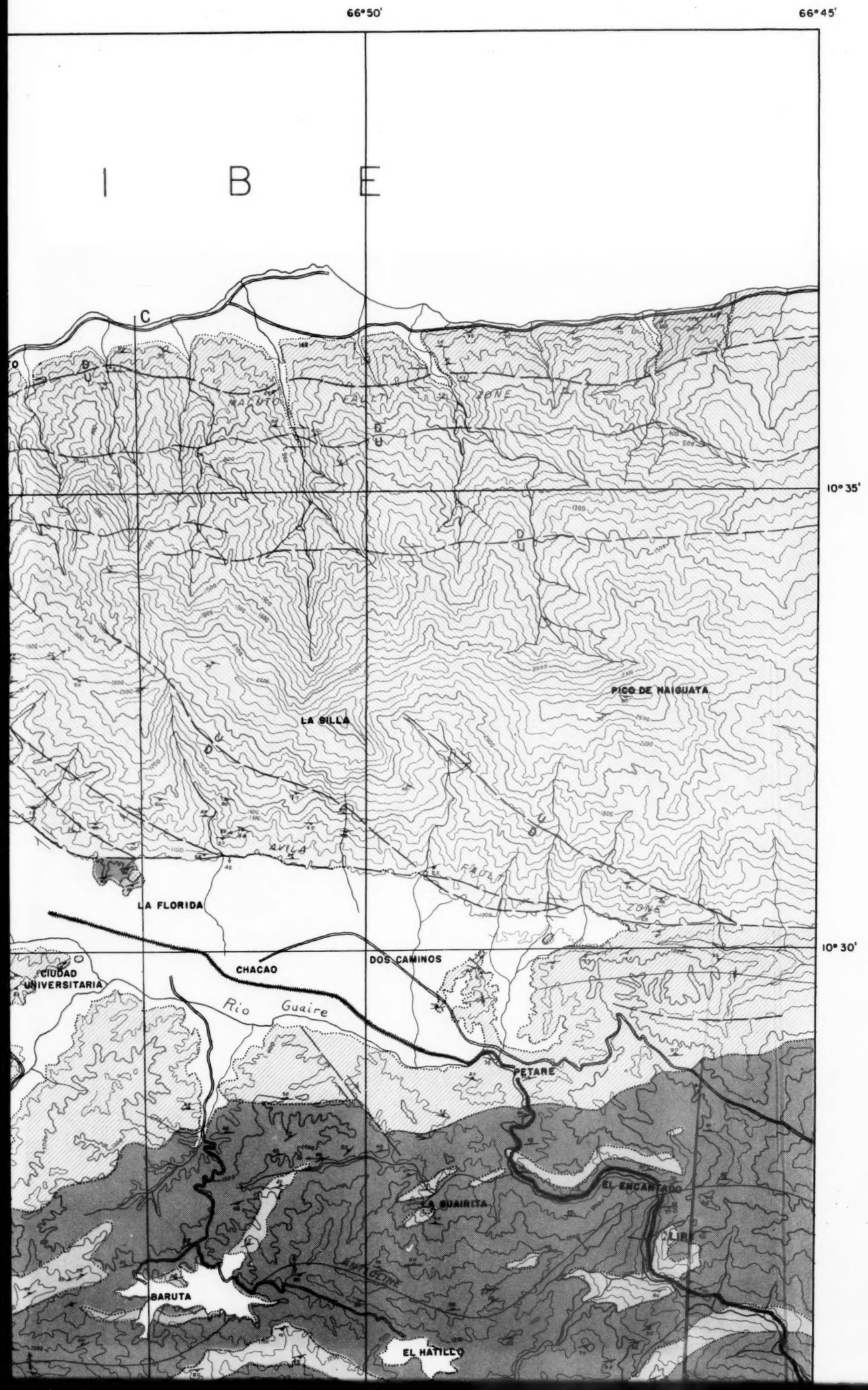
LAS ADJUNTAS

LAS BRISAS

EL HATILLO

LOS DECHOSOS

CARAC



GEOLOGIC MAP

OF THE

CARACAS-LOS TEQUES REGION, VENEZUELA

SCALE 1 : 75, 000

0 1 2 3 4 5

KILOMETERS

CONTOUR INTERVAL 100 METERS

BASE MAP BY CARTOGRAFIA NACIONAL, M. O. P., VENEZUELA

NORTHERN HALF, GEOLOGY BY G. DENGO

SOUTHERN HALF, GEOLOGY BY R. SMITH

Sponsored by Direccion de Geologia,
A. Schwarck Anglade, Director,
Ministerio de Minas e Hidrocarburos,
Estados Unidos de Venezuela

1952

LEGEND:

- ALLUVIUM
- UNDIFFERENTIATED TERTIARY AND QUATERNARY
- CABO BLANCO GROUP
- PARACOTOS FORMATION
- TACAGUA FORMATION
- LAS MERCEDES FM.
  - LOS COLORADOS LIMESTONE FACIES
  - CHARALLAVE CONGLOMERATE MEMBER
- ANTIMANO FORMATION
- LAS BRISAS FM.
  - DOLOMITIC MICROCLINE GNEISS FACIES

10° 30'

10° 25'

10° 20'

EL MANICOMIO

CARACA

PRO-PATRIA

EL PARAISO

EL CEMENTERIO

ANTICLINE

Río Guaire

LA VEGA

ELVALLE

TUNQUITO

Río Macarao

LAS ADJUNTAS

LAS BALSAS

LOS LECHOSOS

LA MARIPOSA

SEBASTOPOL

SAN ANTONIO

SAN PEDRO

LOS TEQUES

CARRIZAL

SAN DIEGO

EL ENCANTO

ROQUERON

A

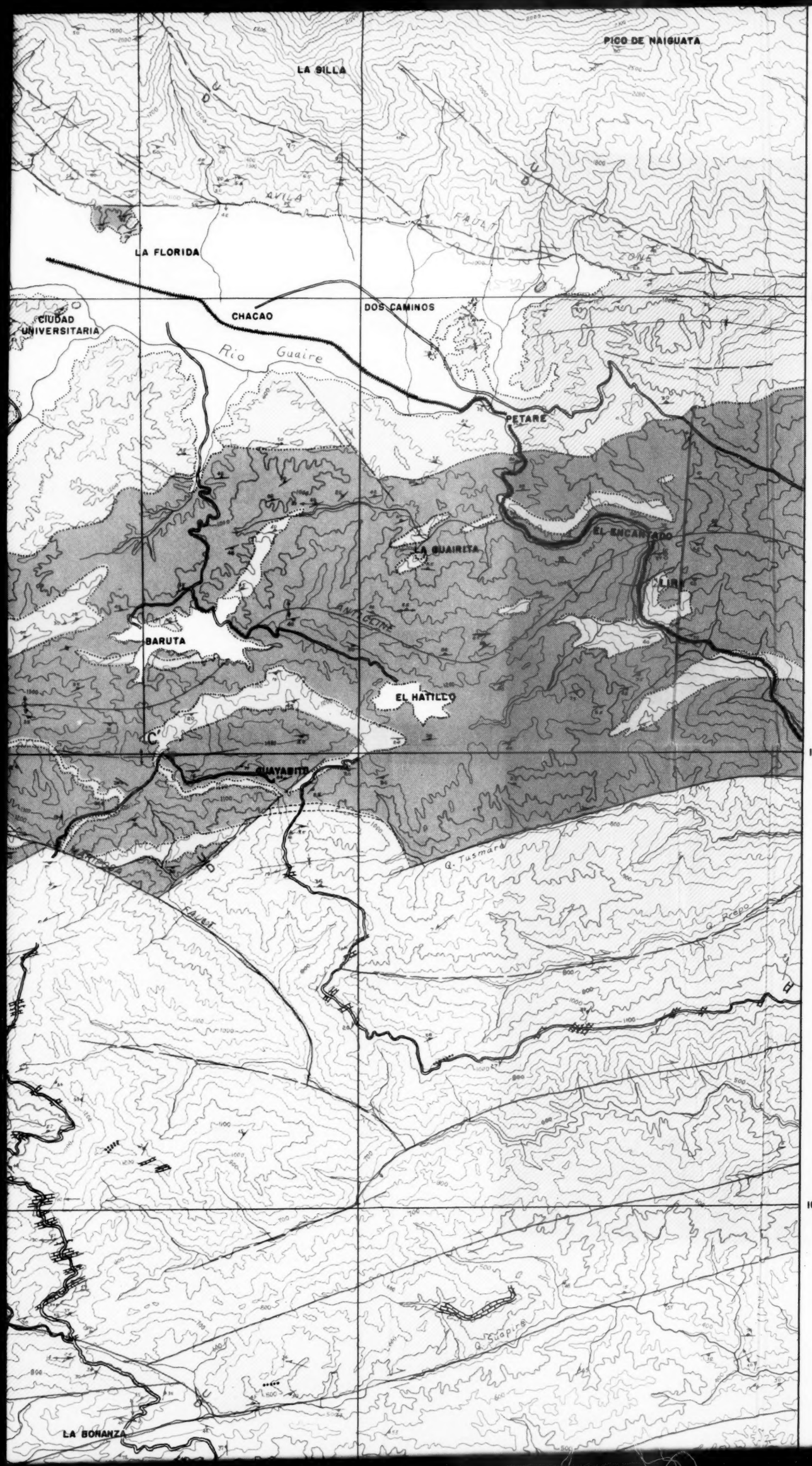
A'

CARACAS

PICO DE NAIGUATA
LA SILLA
EL MANICOMIO
PRO-PATRIA
LA FLORIDA
CHACAO
DOS CAMINOS
EL PARAISO
CIUDAD UNIVERSITARIA
EL CEMENTERIO
LA VEGA
ELVALLE
PETARE
EL ENCANTADO
LA GUAIRITA
LIRI
BARUTA
EL HATILLO
LAS ADJUNTAS
LAS BRISAS
LOS LECHOSOS
LA MARIPOSA
GUAYABITA
SAN ANTONIO
LA CORTADA DEL GUAYABO
CARRIZAL
SAN DIEGO
LOS TEQUES
LA BONANZA
Rio Guaire
Rio Valle
Q. Tusmare
Q. Suapire
AVILA FAULT ZONE
ANTICLINE
A
A'
B
B'
C
C'
D
D'
E
E'

Estados Unidos de Venezuela

1952



LEGEND:

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- ANTIMANO FORMATION
- LAS BRISAS FM.
  - DOLOMITIC MICROCLINE GNEISS FACIES
  - PENA DE MORA GNEISS FACIES
- SEBASTOPOL ORTHOGNEISS BASEMENT
- TIARA VOLCANICS
- AREAS OF ABUNDANT SODA GRANITE DIKES
- HORNBLENDE DIORITE
- ULTRAMAFIC INTRUSIVE ROCKS
- AMPHIBOLITE

DIP AND STRIKE OF BEDDING

DIP AND STRIKE OF FOLIATION
WITH TREND AND PLUNGE OF LINEATION

10° 25'

10° 20'

10° 15'

10° 10'

67° 10'

67° 05'

67° 00'

66° 55'

F

E

D

A'

SEBASTOPOL

LA MARIPOSA

SAN ANTONIO

SAN PEDRO

LAS LAGUNETAS

LOS TEQUES

CARRIZAL

SAN DIEGO

EL ENCANTO

BOQUERON

LOS COLORADOS

Q. Seca

Q. Guayas

GUAYAS

RODRIGUEZ

OCAMPO

Q. Maitana

PARACOTOS

FAULT

Río Tuy

JAIME

THRUST

TACATA

Río Tacata

G'

TIARA

F'

E'

LA MARIPOSA

GUAYABITO

SAN ANTONIO

LA CORTADA DEL GUAYABO

CARRIZAL

LOS TEQUES

SAN DIEGO

LA BONANZA

PARACOTOS

CHARALLAVE

TACATA

CUA

OCUMARE DEL TUY

Río Tuy

FAULT

A

A'

D

D'

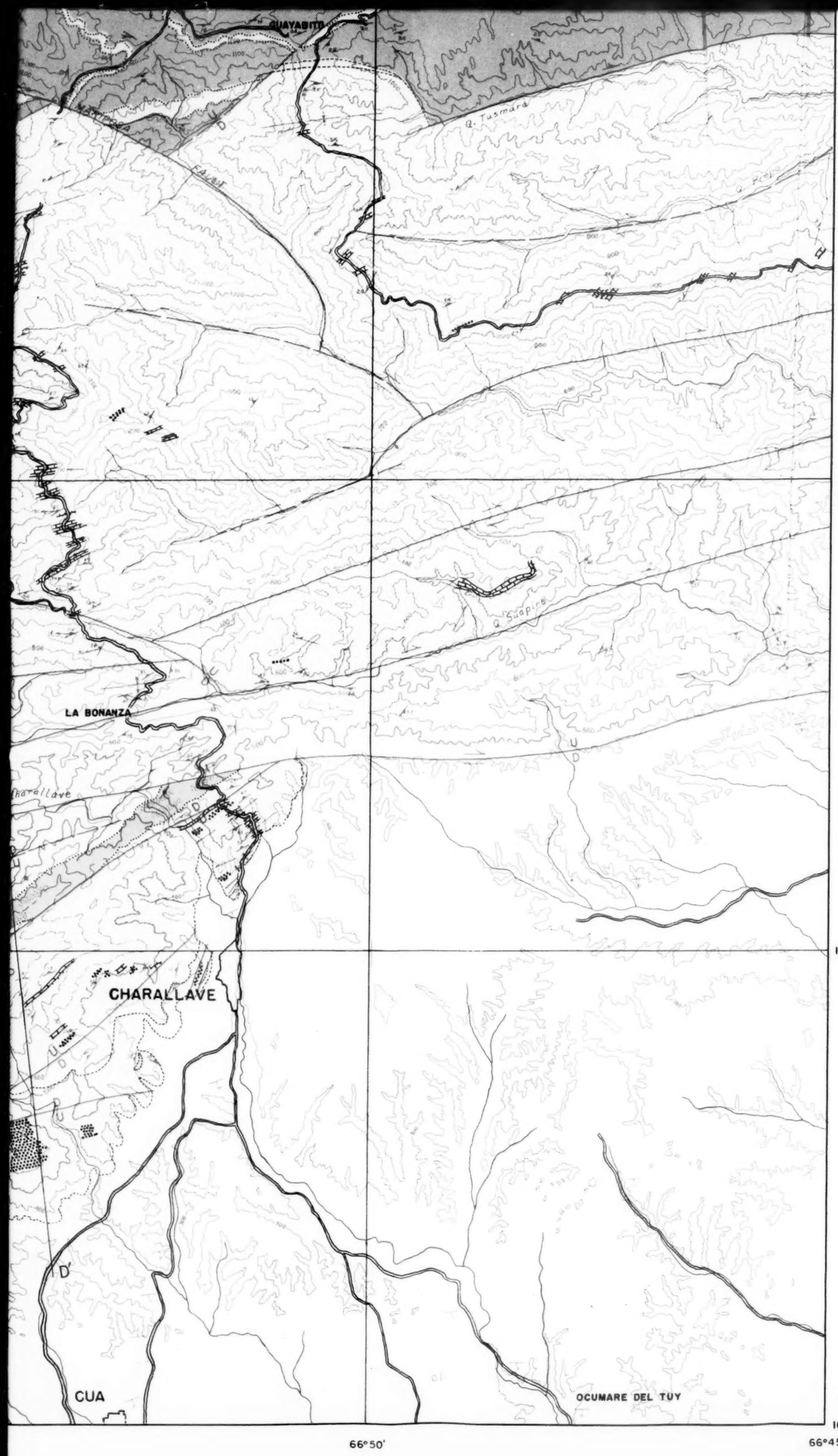
E

E'

67°00'

66°55'

66°50'



PEÑA DE MORA GNEISS FACIES

SEBASTOPOL ORTHOGNEISS BASEMENT

TIARA VOLCANICS

AREAS OF ABUNDANT SODA GRANITE DIKES

HORNBLENDE DIORITE

ULTRAMAFIC INTRUSIVE ROCKS

AMPHIBOLITE

DIP AND STRIKE OF BEDDING

DIP AND STRIKE OF FOLIATION WITH TREND AND PLUNGE OF LINEATION

DIP AND STRIKE OF FOLIATION HORIZONTAL LINEATION

LINEATION

HORIZONTAL FOLIATION

VERTICAL FOLIATION

FAULTS

THRUST FAULTS

FOLD AXES

TREND AND PLUNGE OF MINOR FOLD AXES

QUATERNARY
- Qal — Recent alluvium, shown only where most extensive
- Qcbc / Qcb — Cuerbio basalt, dissected cinder cones and flows
- Qtg — High-level gravels (principally Tuerto gravel)

LATE TERTIARY
- Tsfb / Tsf — Santa Fe formation and interbedded lava flows
- Tma — Abiquiu (?) formation

EARLY TERTIARY
- Tcl — Cieneguilla limburgite
- Tee — Espinaso volcanics
- Teg — Galisteo formation

UPPER CRETACEOUS
- Kmv — Mesaverde formation
- Kmvc — Cano member, Mesaverde formation
- Km — Undivided Mancos shale
  - Km_3 — Upper Mancos shale
  - Km_2 — Middle Mancos shale
  - Km_1 — Lower Mancos shale
    - Km_1 / Kgr / Km_1 / Kd — Subdivisions of lower Mancos shale mapped locally: Greenhorn limestone; Dakota (?) formation

JURASSIC
- J — Entrada sandstone, Wanakah and Morrison formations

TRIASSIC
- Trd — Dockum group

PERMIAN
- Pg — Glorieta sandstone
- Pay — Abo and Yeso formations

PENN.
- Cm — Magdalena group

PRE-CAMBRIAN
- p-€ — Metamorphic and igneous rocks of pre-Cambrian age

- Tb — Basalt dike
- Tp — Non-basaltic dike
- Ti — Non-basaltic intrusions of diverse form

Formation contact

Fault contact, showing relative displacement (D / U)

Fault contact, concealed; shown locally to clarify structural pattern

Attitude of bedding: Normal; Overturned; Horizontal

Locality mentioned in text

× U.S.G.S. 20560 — Fossil locality

City, town or settlement

A ——→ ←—— A' — Trace of structure section

True North

Note: Land lines are shown on the original base maps, but have been omitted here for reasons of clarity.

Scale in Miles: 1/2 0 1 2 3 4

CERROS DEL RIO
RED MT.
TETILLA PEAK
Cieneguilla
MESA NEGRA DE LA BAJADA
La Bajada
MESITA DE JUANA LOPEZ
Rosario Siding
ROSARIO FAULT
Santa Fe
Agua Fria
Santa Fe Creek
SANTA FE AIRPORT
U.S.A AIRPORT
Arroyo Hondo
Cieneguilla Creek
U.S. HIGHWAY No. 85
SANGRE DE CRISTO MTS.
Canyon
La Cieneguilla
Alamo Creek
STATE HIGHWAY No. 10
SANTA FE PLATEAU
CAVALRY BUTTE
Bonanza
KENTUCKY RIDGE
GALISTEO MONOCLINE
Los Cerrillos
Madrid
Rogers
Galisteo Creek
Burley Hill
Kennedy
Galisteo
CRESTON
GLORIETA MESA
SANTO DOMINGO VALLEY
post-Cuerbio basalt
San Felipe
Arroyo Seco
bed of pumice
Arroyo Tonque
ESPINASO RIDGE
Arroyo de las Angeles
Arroyo Pinoncillo
Arroyo de las Tunas
Arroyo de la Vega
Arroyo Alamos Altos
Algodones
A.T.S.F. Railroad
RIO GRANDE
Alamos Altos Ranch
Tonque
HUERFANO BUTTE
Tejon
Arroyo las Huertas
Arroyo Chimal
SAN FRANCISCO FAULT
BERNALILLO FAULT
Placitas
Arroyo Agostura
Arroyo Galisteo
Arroyo Jara
Arroyo San Marcos
Arroyo San Jaspe
Dolores
ORTIZ MOUNTAINS
NOT MAPPED
CERRO PELON
CANYADA LAS JOLLAS
Omero
Additional faults in this area not mapped
Additional faults in this area not mapped
concealed by

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1 ½ 0 1 2 3 4
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Cieneguilla
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bed of pumice
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Arroyo Angostura
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SAN FRANCISCO FAULT
BERNALILLO FAULT
Placitas
Tejon
Tonque
HUERFANO BUTTE
ESPINASO RIDGE
Arroyo de la Vega
Arroyo de las Angeles
Arroyo Pinonchito
Arroyo Chamita
GALISTEO MONOCLINE
Rogers
Madrid
Los Cerrillos
Dolores
Burley Hill
Kennedy
Galisteo
Creston
Arroyo Jara
Arroyo San Marcos
Arroyo Galisteo
Arroyo Chupadero
CERRO PELON
CANYADA LAS JOLLAS
ORTIZ MOUNTAINS
NOT MAPPED
GLORIETA MESA
Additional faults in this area not mapped
Additional faults in this area not mapped
concealed by

GEOLOGIC MAP OF THE GALISTEO-TONQUE AREA
NEW MEXICO

Geology by C.E. Stearns, 1939-1946, assisted by A.F. Marston, J.K. Grunig, N.O. Parker, and I.S. Fisher.

Base map from planimetric drainage maps compiled by the Soil Conservation Service from aerial photographs.

Explanation

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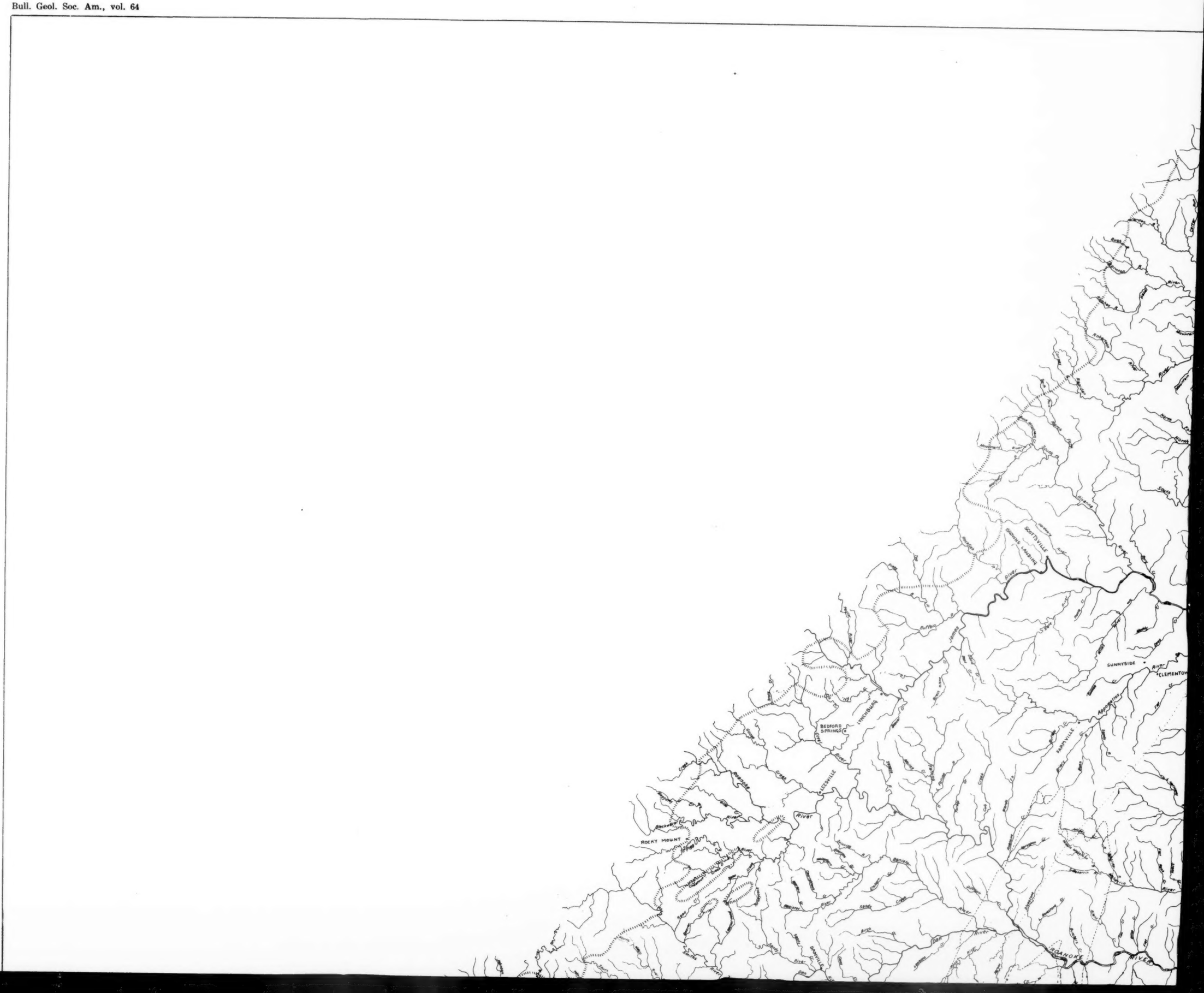
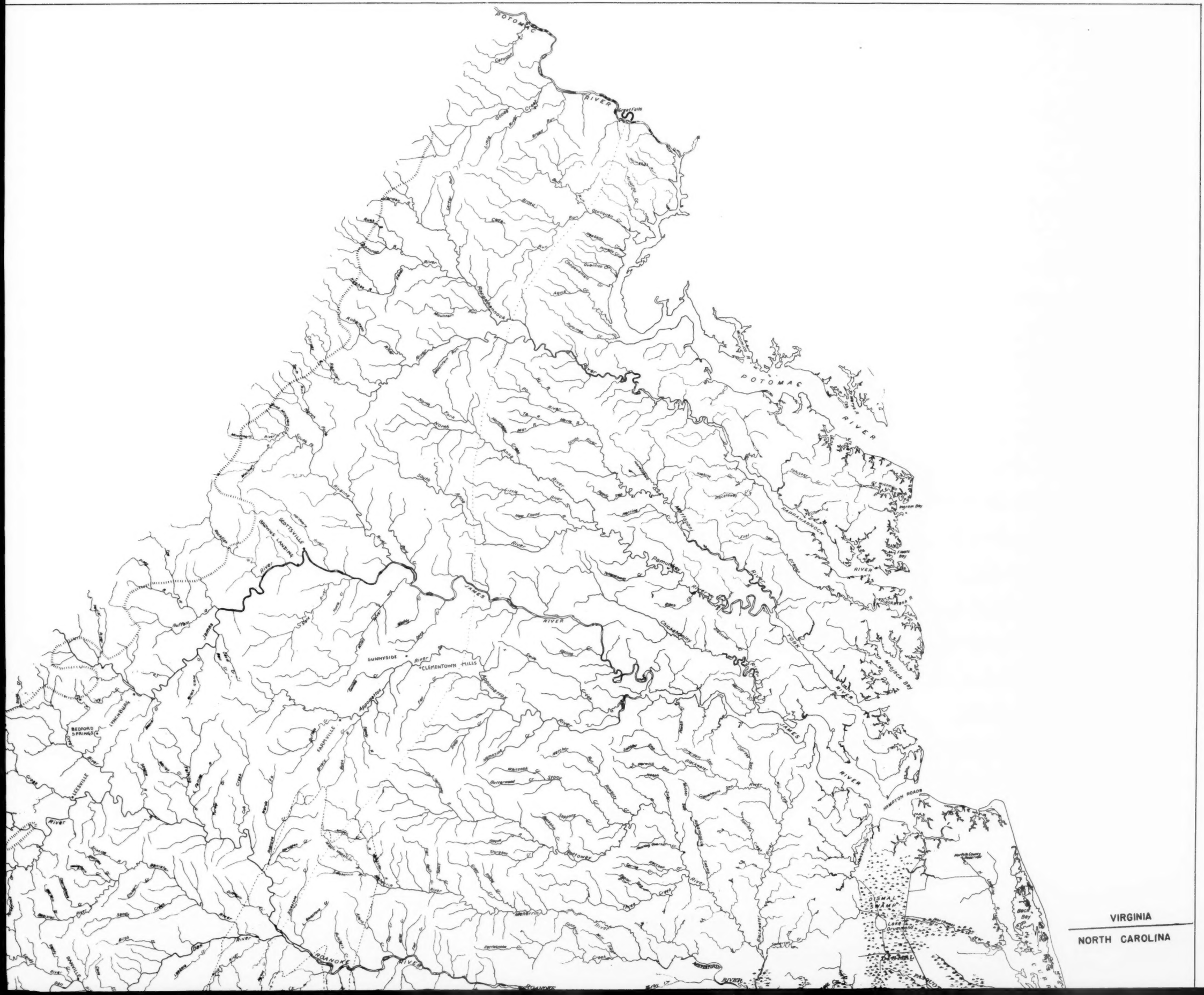
DET·PUB·LIB


SCOTTSVILLE
BROWNS LANDING
SUNNYSIDE
CLEMENTON
BEDFORD SPRINGS
LYNCHBURG
FARMVILLE
ROCKY MOUNT
DANVILLE
ROANOKE
RIVER


POTOMAC RIVER
Great Falls
Rappahannock River
POTOMAC RIVER
Fredericksburg
RAPPAHANNOCK RIVER
YORK RIVER
MOBJACK BAY
JAMES RIVER
SUNNYSIDE
CLEMENTOWN
SCOTTSVILLE
BROWNS LANDING
BEDFORD SPRINGS
LYNCHBURG
FARMVILLE
Appomattox River
JAMES RIVER
HAMPTON ROADS
DISMAL SWAMP
Back Bay
ROANOKE RIVER
VIRGINIA
NORTH CAROLINA

VIRGINIA

NORTH CAROLINA

N

LEGEND

S.....S Former edge of Coastal Plain as suggested by drainage pattern.

--------- Limits of Carolina Slates.

Continuous line of hachures along northwest side of map represents foot of Blue Ridge Scarp.

0 10 20 30

Miles

DRAINAGE MAP OF PARTS OF NORTH CAROLINA AND VIRGINIA, SHOWING PIEDMONT AND ADJA

(After U.S. Geological Survey.)

DET·PUB·LIB

VIRGINIA, SHOWING PIEDMONT AND ADJACENT PARTS OF COASTAL PLAIN AND APPALACHIAN MOUNTAINS

(After U.S. Geological Survey.)